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New Data on the Lower Cambrian Biostratigraphy of the Eastern Sayan

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Abstract—Geological mapping of the Eastern Sayan in the late 1980s–early 1990s proved the occurrences of stable assemblages of archaeocyathids and trilobites correlatable with those from horizons of the Altai–Sayan foldbelt (except for the Ust'-Kundat Horizon). Archaeocyathids and trilobites provide reliable correlations of the Kiya, Kameshki, Sanashtykgol, and Obruchev horizons both within the Eastern Sayan area and between the neighboring areas of the Altai–Sayan foldbelt. The small-shelled fauna associated with archaeocyathids in the Lower Cambrian sequences allow us to accurately date basal Cambrian beds barren of archaeocyathids near the Vendian boundary. Of interest are the occurrences of small-shelled forms typical of the Tommotian Stage in the Natal'evka, Kiya, Kameshki, and Sanashtykgol horizons.

Key words: *Eastern Sayan, Lower Cambrian, biostratigraphy, archaeocyathids, small-shelled fauna.*

INTRODUCTION

Fossils still serve as the main tool for subdivision and correlation of the Lower Cambrian bio- and chronostratigraphic units. At present, much less sequences received thorough paleontological study in geosynclinal areas than in platforms, e.g., no recent publications on fossil assemblages of the Eastern Sayan are available. To elaborate legends of large- and middle-scaled geological maps, we investigated the Lower Cambrian stratotype areas of the Eastern Sayan. For the first time, the earliest small-shelled faunas were found together with abundant archaeocyathids and renalcids in the Lower Cambrian deposits of the Torgashino, Sretenka, Balakhtison, and Chibizhek areas (Figs. 1 and 2). The archaeocyathid assemblages are well correlated with those of the Natal'evka, Kiya, Kameshki, Sanashtykgol, and Obruchev horizons of the Altai–Sayan foldbelt. The assemblages of small-shelled forms are of particular interest for recognizing horizons of the Eastern Sayan correlatable with the Tommotian Stage.

THE TORGASHINO STRATOTYPE AREA

The Torgashino Ridge is located in the extreme northwestern part of the Eastern Sayan to the south of the town of Krasnoyarsk (Fig. 1-I). Sections of its southern and northern slopes are well known as the reference sections of the carbonate Cambrian of the whole Eastern Sayan, and the section in the right-hand side of the Bazaikha River opposite the Kaltat River mouth is accepted as the stratotype of the Bazaikha Superhorizon. According to the unified stratigraphic scheme for the Altai–Sayan foldbelt, this superhorizon consists of the Natal'evka and Kiya horizons and corresponds to the lower part of the Atdabanian Stage of the Siberian platform (*Resheniya ...*, 1983). The Cambrian biostratigraphy of the Torgashino Ridge was most thoroughly studied by Zadorozhnaya, Zhuravleva, Kashina, Repina, Rozanov, and Missarzhevskii (Repina, 1960; Repina *et al.*, 1964; Rozanov and Missarzhevskii, 1966; Zadorozhnaya *et al.*, 1972; Kashina and Yankauskas, 1973; Osadchaya *et al.*, 1979). Their data have been recently summarized by Astashkin *et al.* (1995). Our investigations of 1989–1993 supplement these materials.

The almost continuous section of the Kaltat Formation underlying the Torgashino Formation and previously correlated with the Ust'-Kundat Horizon of the Altai–Sayan foldbelt and the Ungut Formation of the Mana trough (Khomentovskii *et al.*, 1960, 1978; Koptev, 1961a, 1961b, 1962; Predtechenskii, 1967; Rozanov *et al.*, 1969) is described below. The section is located in the right bank of the Bazaikha River, 600 m downstream from the Kaltat River mouth. The following beds are exposed here beginning from the level of 6 m above the water surface upward:

The Kaltat Formation (overall thickness is 30 m):

(1) dark-gray wavy bedded clayey limestones (10 m);

(2) greenish gray fine- to medium-grained micaceous calcareous sandstones (2 m); the bed yielded small-shelled fauna *Hyolithellus* sp., *H. tenuis* Miss., *Torellella* sp., and *Siphogonuchites* sp. in association with renalcids *Epiphyton* sp., sponge spicules of Protospongiidae, inarticulate brachiopods, and archaeocyathid fragments (identifications by A.V. Efimova);

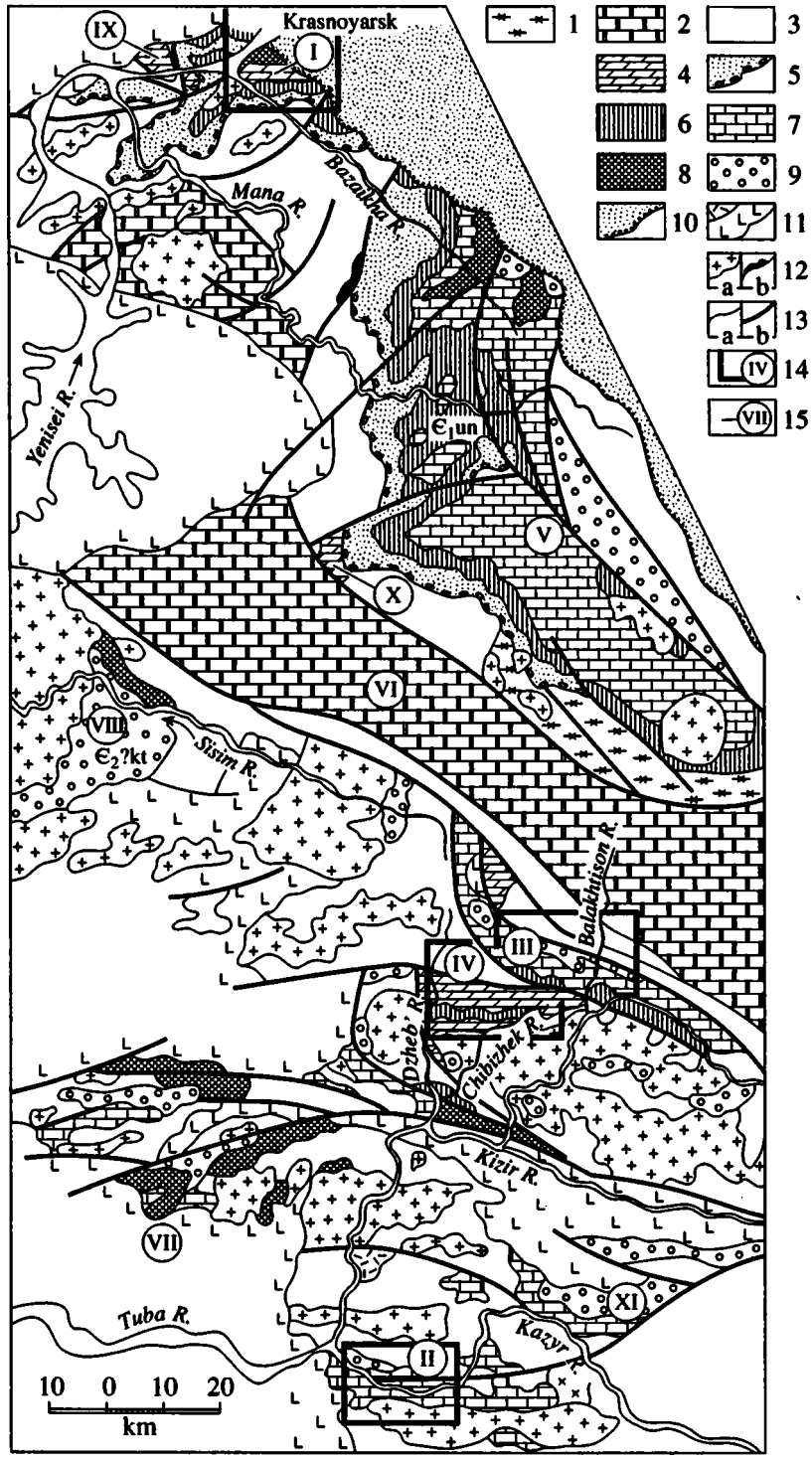

Fig. 1. Schematic map of Vendian-Cambrian rocks in the Eastern Sayan: (1) Archaen gray greisses; (2) Proterozoic graphitic marbles and crystalline schists; (3) Riphean Urma, Mana, and Bakhtino formations; (4) Vendian carbonates of the Ovsyanka, Angaloi, and lower Chibizhek formations; (5–6) Vendian–Cambrian transgresive deposits: (5) graywackes with limestone interbeds of the Anastas'ino Formation, (6) massive and flaggy dolomitic limestones of the Ungut and upper Chibizhek formations; (7) reefal and bedded carbonates of the Atdabanian and Toyonian stages of the lower Cambrian; (8) carbonate-terrigenous deposits of the Amgian Stage of the Middle Cambrian; (9) terrigenous-volcanic regressive deposits of the Middle (?) Cambrian; (10) sedimentary and (11) volcanic deposits in Devonian depressions; (12) granitoid (a) and ultramafic (b) intrusions; (13) geological boundaries (a) and tectonic contacts (b); (14) boundaries and numbers of stratotype areas: I, Torgashino Ridge; II, Sretenka; III, Balakhtison; IV, Chibizhek area; (15) other areas of significance: V, Mana depression; VI, Kameshki Valley; VII, Derba anticlinorium; VIII, stratotype areas of the Middle Cambrian Lodochnoe and Middle (?) Cambrian Kotel formations; IX, Ovsyanka Formation stratotype; X, Angaloi Formation stratotype; XI, Cheremshanka Formation stratotype.

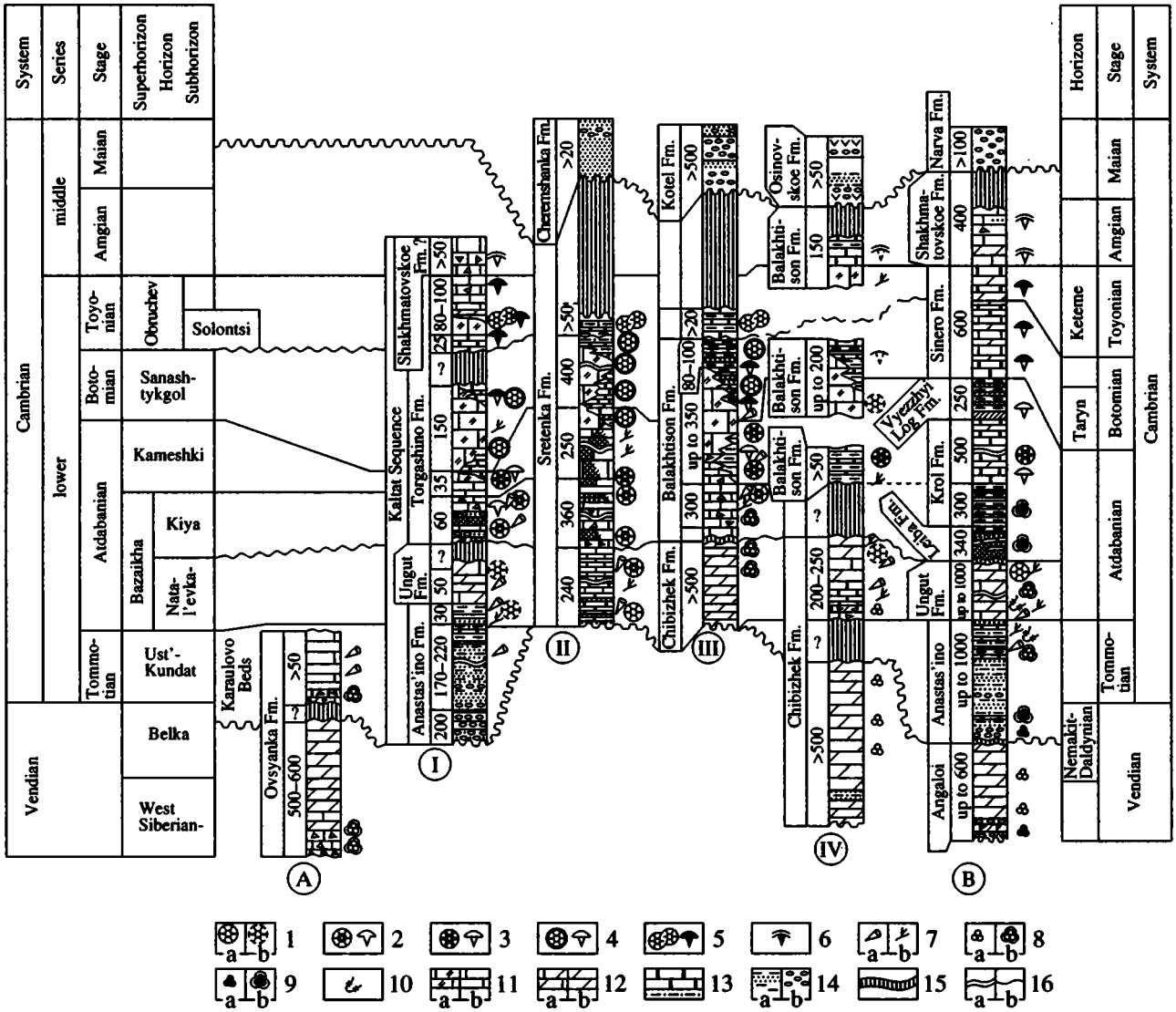

Fig. 2. Correlation scheme of Vendian–Cambrian deposits in the western slope of the Eastern Sayan.
(1–10) fossil assemblages: (1) abundant (a) and isolated (b) archaeocyathids of the Natal'evka Horizon; (2) archaeocyathids and trilobites of the Kiya Horizon; (3) archaeocyathids and trilobites of the Kameshki Horizon; (4) archaeocyathids and trilobites of the Sanashtykgol Horizon; (5) archaeocyathids and trilobites of the Obruchev Horizon; (6) trilobites of the Amgian Stage; (7) small-shelled fauna (a) and (b); (8) Vendian (a) and Cambrian (b) microphytoliths; (9) Vendian (a) and Cambrian (b) phytoderivates; (10) cribricyathids; (11) light massive reefal (a) and red (b) limestones; (12) dolomitized limestones (a) and dolomites (b); (13) dark to black bedded limestones and calcareous-clayey siltstones; (14) sandstones or siltstones (a) or conglomerates (b); (15) sedimentation breaks; (16) reduced thickness (a) and unconformity (b). Stratotype areas: I, Torgashino, II, Sretenka, III, Balakhtison, IV, Chibizhek; (A) the Ovsyanka Formation stratotype (Fig. 1-IX); (B) the Mana depression (Fig. 1-V).

(3) alternating greenish gray sandstones, siltstones, and marls with rare interbeds of lenticularly bedded limestones (5 m);

(4) light and bluish gray wavy bedded limestones, whose individual beds increase upward in thickness from 3 to 5–6 cm (6 m);

(5) alternating sandstones and calcareous siltstones with pale interbeds of conglomerate-like (turbidites) limestone (2 m);

(6) poorly exposed interval with debris of red thinly bedded calcareous siltstones (5 m).

The Torgashino Formation, the Bazaikha Member (overall thickness is 60 m):

(7) red calcareous siltstones (2 m);

(8) poorly exposed interval with debris and deluvium of red calcareous breccia of gravelstones and clastic limestones (20 m);

(9) pinkish brecciated limestones with interbeds of pink calcareous gravelstones and breccias (14 m); the

limestone beds contain archaeocyathids *Archaeolynthus sibiricus* (Toll), *Tumuliolynthus tubexternus* (Vol.), *Capsulocyathus subcallosus* Zhur., *Nochoroicyathus* sp., *Orbicyathus* sp., *Tumulocyathus* sp., *Kijacyathus* sp., *Plicocyathus mirus* (Osad.), *Alataucyathus* sp., *Cambrocyathellus* sp., and *Dictyosycon radiatus* (Zhur.) (identifications by I.T. Zhuravleva) in association with trilobites *Resimopsis* sp., and *R. mariinica* Rep. (identifications by Repina); gravelstones yielded small-shelled fauna: *Tommotia plana* (Miss.), *Rhombocorniculum cancellatum* (Cobb.), Chancelloriidae gen. et sp. indet., *Conotheca circumflexa* Miss., *Hyolithellus* sp., and *Cambrocassis* cf. *verrucatus* Miss. (identifications by Efimova);

(10) pinkish gray brecciated limestones with gravelstone–breccia interbeds;

Archaeocyathids from the last bed are *Archaeolynthus* sp., *Gordonicyathus subhowelli* (Osad.), *Taylorcyathus subtersiensis* (Vol.), *Tumulocyathus* sp., *Baikalocyathus* sp., *Erismacoscinus* sp., *Kijacyathus* sp., *Retecoscinus retetabulae* (Vol.), and *Cambrocyathellus* sp. (identifications by I.T. Zhuravleva).

Opposite the Kaltata River mouth (Mt. Pioner locality), the small-shelled fauna of *Hyolithellus* sp., Chancelloriidae gen. et sp. indet., *Sachites* sp., Helcionellaceae (?) gen. et sp. indet. was found together with sponge spicules of Protospongiidae (identifications by Efimova), and renalcids of genera *Epiphyton, Renalcis, Proaulopora, Bija, Razumovskia, Subtifloria, Batinevia*, and *Botomaella* (identifications by Popov, 1991 and Luchinina, 1986) in the exposed dark-gray flaggy sandy limestones of the middle Kaltat Formation.

The terrigenous-carbonate Kaltal deposits and the underlying graywackes and conglomerates of the Anastas'ino Formation occur in a synclinal structure (Fig. 1). The rhythmical interbedding, turbidites, and fossil assemblage make the Kaltat Formation very similar to the lower Ungut and upper Anastas'ino formations of the Mana depression. The Kaltat deposits are overlain by the Bazaikha Member. The predominant archaeothyathid assemblage from red deposits of the Bazaikha Member is of the Kiya age. Upsection, the Torgashino Formation contains three successive (Kameshki, Sanashtykgol, and Obruchev) archaeocyathid assemblages. This subdivision is in agreement with the trilobite distribution pattern (Dyatlova, 1990).

THE SRETENKA STRATOTYPE AREA

In the extreme southwestern part of the Eastern Sayan, the Lower Cambrian deposits are exposed in lower courses of the Kazyra River between the Tayaty and Sretenka villages (Fig. 1, II).

The most comprehensive study and description of this section were done by D.I. Musatov, V.N. Nemirovskaya, E.V. Shirokova, and I.T. Zhuravleva as far back as 1961. The materials were briefly summarized by Osadchaya *et al.* (1979). Only in the mid-1980s, geologists of the Minusinsk Expedition obtained some new geological and paleontological data in the course of large-scaled geological mapping.

In 1986–1987, a large paleontological collection from Sretenka, Balakhtison, and Chibizhek sections was studied by paleontologists of PG "Krasnoyarskgeologos"emka", including Dyatlova (archaeocyathids), Shipitsyn (microphytoliths and renalcids), and Efimova (small-shelled fauna). In 1986, geologists of the Minusinsk Expedition headed by V.N. Vorob'ev attributed the reefogenic-carbonate Lower Cambrian deposits to the Sretenka Formation resting on basic and intermediate volcanites.

The section begins in the right bank of the Kasyra River opposite the Tayaty Village, where andesite–basaltic volcanites are overlain by the following beds (upsection from the base):

(1) clastic brecciated limestones alternating with gray massive and charcoal gray bedded limestones (100 m) with archaeocyathids *Archaeolynthus sibiricus* (Toll), *Nochoroicyathus* sp., *Tumuliolynthus* sp., *Tumulocyathus* sp., and *Dictyocyathus* sp. associated with renalcids of genera *Epiphyton, Proaulopora, Botominella*, and *Girvanella*, and small-shelled fauna *Hyolithellus* sp. and *Torellella* sp.;

(2) light-gray bedded limestones with interbeds of dark-gray to black limestones; the unit includes taphoherms, no bioherms (140 m), and archaeocyathids: *Archaeolynthus sibiricus* (Toll), *Tumuliolynthus* sp., *Fransuasaecyathus subtumulatus* Zhur., *Nochoroicyathus arteintervallum* (Vol.), *Rotundocyathus proskurjakowi* (Vol.), *Plicocyathus* ex gr. *admirabilis* (Vol.), and *Pretiosocyathus* sp. coexisting with renalcids of genera *Epiphyton, Razumovskia*, and *Renalcis*; small-shelled fauna of *Hyolithellus* sp., *H. tenuis* Miss., *H.* cf. *isiticus* Miss., *H.* ex gr. *isiticus* Miss., *Torellella* sp., *Sunnaginia* sp., *S.* cf. *imbricata* Miss., *Tommotia* sp., *Coleolella* sp., *Coleolus* sp., *C.* cf. *trigonus* Sys., and Chancelloriidae gen. et sp. indet., and also with brachiopods *Kutorgina* sp.; according to archaeocyathids, the beds are attributed to the Natal'evka Horizon of the Atdabanian Stage, which is 240 m thick in the section;

(3) light pinkish, massive to flaggy limestones, frequently clastic and detrital (up to 200 m); the rocks bear archaeocyathids of the Kiya Horizons of the Atdabanian Stage: *Nochoroicyathus spinosus* (Vol.), *N. fragilis* Osad., *N. mariinskii* Zhur., *N.* ex gr. *arteintervallum* (Vol.), *Tumulocyathus pustulatus* Vol., *T.* ex gr. *pustulatus* Vol., *Urcyathella tercyathoides* Zhur., *Orbicyathus* sp., *Inessocyathus* sp., *Archaeopharetra* sp., *Rotundocyathus* sp., *Dictyocyathus* sp. and renalcids of genera *Epiphyton, Gordonophyton, Razumovskia, Renalcis*, and *Girvanella*;

(4) gray massive–flaggy limestones grading upsection into light variegated clastic limestones; the rocks are intercalated with thin interbeds of red gravelly sandstones (160 m) and bear archaeocyathids of the Kiya Horizon: *Archaeolynthus sibiricus* (Toll), *Tumuliolyn-*

thus sp., *Orbicyathus* ex gr. *kundatus* Vor., *Rotundocyathus ijizkii* (Toll), *Tumulocyathus pustulatus* Vol., *Nochoroicyathus* cf. *fragilis* Osad., *Alataucyathus* sp., *Mennericyathus shoriensis* (Roz.), and *Dictyocyathus* sp. associated with renalcids of genera *Epiphyton, Gordonophyton, Razumovskia, Girvanella, Botominella, Batinevia, Proaulopora*, and *Kordephyton*;

(5) pale limestones with red sandstone interbeds at the base and gray flaggy limestones with chert interbeds at the top (150 m); medium-sized renalcid bioherms are predominant in the unit also containing renalcid–archaeocyathid bioherms in its lower part; the archaeocyathid assemblage is transitional between those from the Kiya and Kameshki horizons (the Atdabanian Stage) and includes *Nochoroicyathus* sp., *N.* aff. *howelli* (Vol.), *Rotundocyathus* cf. *ijizkii* (Toll), *Tumulocyathus pustulatus* Vol., *Plicocyathus volgaris* (Osad.), *Leptosocyathus* ex gr. *regularis* Vol., *Pretiosocyathus* sp., *Arturocyathus* sp., *Coscinocyathella* sp., and *Dictyocyathus* sp.; coeval renalcids represent genera *Epiphyton, Gordonophyton, Proaulopora, Razumovskia*, and *Girvanella*;

(6) pale massive renalcid limestones (up to 50 m) with species of genera *Epiphyton, Gordonophyton, Kordephyton, Razumovskia, Renalcis, Girvanella, Botominella*, and *Proaulopora*, which are thought to be characteristic of the Kameshki Horizon;

(7) light archaeocyathid–renalcid limestones with large bioherms (up to 100 m); the rocks yielded archaeocyathids of the Sanashtykgol Horizon of the Botomian Stage: *Coscinocyathella* sp., *Tenneticyathus* sp., *Lenocyathus* sp., *Aptocyathus* sp., and *Dictyocyathus* sp.; associated renalcids are from genera *Epiphyton* and *Proaulopora*;

(8) light massive limestones with algal–archaeocyathid bioherms (up to 100 m) with archaeocyathids of the Sanashtykgol Horizon, such as *Gordonicyathus howelli* (Vol.), *Archaeolynthus* sp., *Erismacoscinum* ex gr. *macrospinosus* (Zhur.), *Capsulocyathus irregularis* (Zhur.), *Fransuasaecyathus subtumulatus* Zhur., *Carinacyathus* ex gr. *mirabilis* (Zhur.), *Clathricoscinus* sp., and *Anthomorpha* ex gr. *rackovskii* (Vol.), and also with renalcids of genera *Epiphyton, Renalcis, Kordephyton, Girvanella*, and *Tarthinia*;

(9) gray massive limestones (up to 50 m) bearing archaeocyathids of the middle-upper part of the Sanashtykgol Horizon, namely *Capsulocyathus irregularis* (Zhur.), *Pycnoidocyathus* ex gr. *erbiensis* (Zhur.), *Irinaecyathus ratus* (Vol.), *Ladaecyathus* ex gr. *limbatus* Zhur., *Kazyricyathus* sp., *Syringocyathus* sp., *Gordonicyathus* sp., *Clathricoscinus* sp., *Leptosocyathus* sp., and *Anthomorpha* sp., which associate with renalcids of genera *Epiphyton, Renalcis, Razumovskia, Kordephyton, Tubomorphophyton, Tarthinia*, and *Girvanella.*

The last unit is unconformably overlain by variegated polymictic sandstones and gravelstones of the Cheremshanka Formation of the Middle (?) Cambrian.

Dark-gray to black carbonate shales overlying basaltic andesites in the area are referred to as the Kolpa Formation of the Lower Cambrian, which occurs below the Sretenka Formation according to mapping results of the Minusinsk Expedition. Data on archaeocyathids and renalcids from black limestone interbeds suggest that these deposits are not older than the upper Botomian Stage (Fig. 2).

THE BALAKHTISON STRATOTYPE AREA

The third Lower Cambrian section is located in the middle part of the western slope of the Eastern Sayan at the Balakhtison River (Figs. 1, III and 3b).

The Lower Cambrian Balakhtison Formation was established under this name by F.Ya. Pan in 1951 (Khomentovskii *et al.*, 1960). At that time, it was considered as a unit of sandy–clayey and carbonate shales with dark-gray limestone interbeds and reefal buildups containing Early Cambrian archaeocyathids and trilobites. The overlying conglomerates were assigned by Pan to the Devonian. Biostratigraphy of the section was described in a single publication (Khomentovskii *et al.*, 1960). Later, at the time of middle-scaled geological mapping, A.D. Shelkovnikov in 1962 and V.M. Chairkin in 1968 identified the Lower Cambrian deposits of the area as the Kolpa (predominantly shales) and Balakhtison (reefal limestones) formations. Up to the present, the stratigraphic succession and scope of the Lower Cambrian deposits of the Balakhtison section remain disputable. The fossils collected by geologists of the Minusinsk Expedition in 1986–1987 and the results of our field investigations in 1988 proved the Pan's version as the most acceptable.

In spite of numerous tectonic faults and heavy folding, the Balakhtison section demonstrates the formation base and the contact with overlying deposits (Fig. 3b).

On the right bank of the Balakhtison River, from the south to the north of the Kolpa River mouth, the section is composed of the following deposits.

The Chibizhek Formation:

(1) light-gray massive and flaggy dolomitized limestones containing large oncolites and microphytoliths *Osagia* ex gr. *grandis* Z. Zhur., *O. lamellata* Korde, *Ambigolamellatus* Z. Zhur., and *Vesicularites angularis* Reitl. (up to 800 m).

The Balakhtison Formation:

(2) alternation of black carbonate–clayey shales, siltstones, and sandstones with black limestone interbeds (up to 500 m); in limestones Repina found trilobites *Redlichina* sp. characteristic of the Sanashtykgol Horizon of the Botomian Stage (Khomentovskii *et al.*, 1960);

(3) rhythmically bedded, fine-banded limestones with interbeds of sedimentary breccias (up to 200 m); breccias yielded microphytoliths *Osagia* cf. *senta* Z. Zhur., *O.* ex gr. *grandis* Z. Zhur., *O. kolbinica* Yaksch., *Ambigolamellatus horridus* Z. Zhur., *Vesicularites*

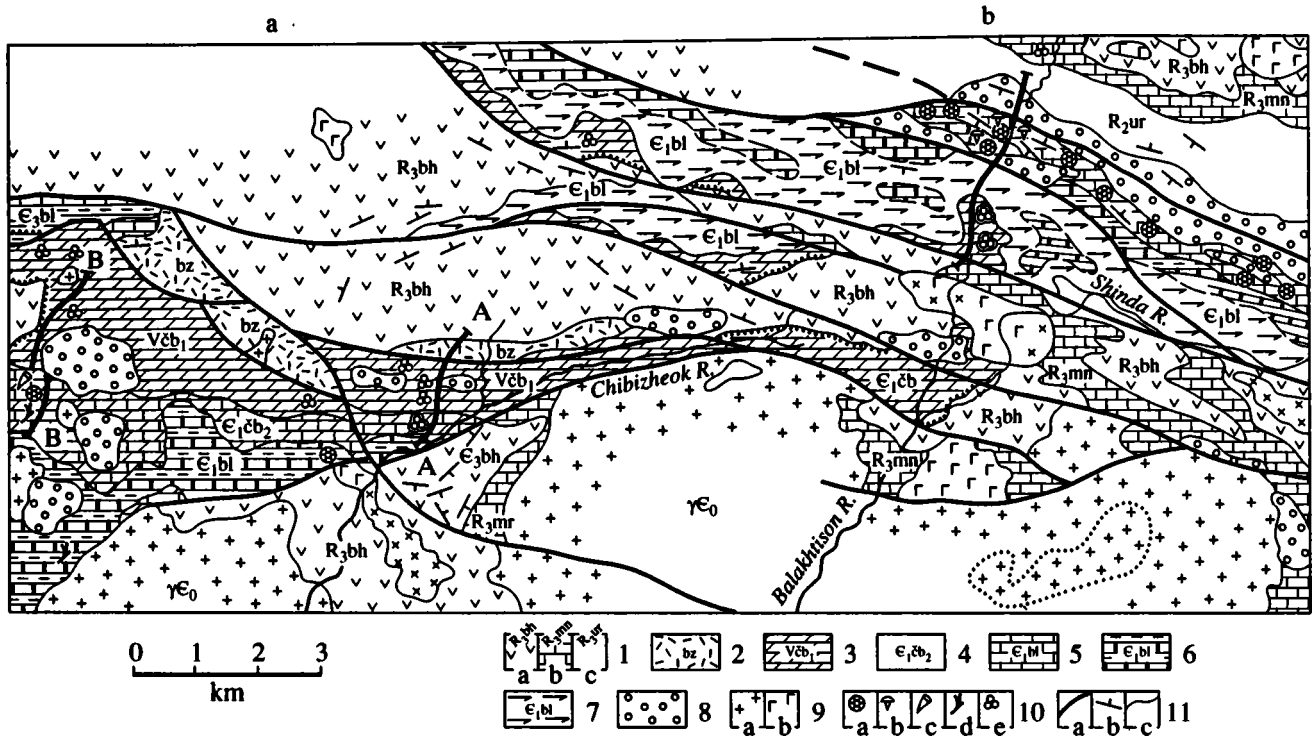

Fig. 3. Schematic map of the Chibizhek (a) and Balakhtison (b) areas.
(1) Riphean volcanites of the Bakhta Formation (a), limestones of the Mana Formation (b), and shales of the Urma Formation (c); (2) tufogenic-sedimentary nameless sequence; (3–4) dolomites with the Yudoma microphytolite assemblage (3) and dolomitic limestones with Cambrian fossils (4) of the Vendian–Lower Cambrian Chibizhek Formation; (5–7) light-colored massive reefal limestones (5), dark-gray bedded limestones intercalated with calcareous-clayey shales (6), and black, gray, or variegated clayey-carbonate shales intercalated with less frequent siltstones and sandstones (7) of the Atdabanian–Botomian Balakhtison Formation; (8) gray and variegated conglomerates, gritstones, and sandstones of the Middle (?) Cambrian Kotel Formation; (9) acid to intermediate (a) and basic (b) intrusions; (10) localities of archaeocyathids (a), trilobites (b), small-shelled fauna (c), renalcids (d), and microphytolites (e); (11) symbols of faults (a), bed attitude (b), and geological boundaries (c).

artus Yaksch., *Vermiculites tortuosus* Reitl., and *V. angularis* Reitl;

(4) dark thinly stripped limestones with interbeds of oolitic limestones (250 m).

Succession of the rhythmic members are difficult to establish in this part of the Balakhtison section, which is disharmonically folded near the tectonic thrust. The third bed with interbeds of sedimentary breccias seems to be lowermost.

After the tectonic fault running along the valley of sublatitudinal direction, there are exposed beds from the south to the north;

(5) carbonate shales and clayey limestones crown by light-gray massive limestones (350 m); rocks of the last type contain archaeocyathids and other taxa presumably characteristic of the Kiya Horizon of the Atdabanian Stage: *Nochoroicyathus* sp., *Pretiosocyathus* cf. *subtilis* Roz., *Loculicyathus membranivestites* Vol., *Dictyocyathus* sp., calcified cyanobacteria of genera *Batinevia, Botomaella, Subtifloria, Proaulopora,* small-shelled *Hyolithellus* sp., and inarticulate brachiopods;

(6) clayey stripped limestones and interbeds of red-gray pelitomorphic limestones (50 m); the rocks yielded archaeocyathids of the Kiya Horizon: *Archaeolynthus* sp., *Nochoroicyathus* sp., *Arturocyathus* cf. *borisovi* Roz., *Nalivkinicyathus* cf. *cyroflexus* (Bojar. et Osad.), *Erismacoscinus* sp., *Plicocyathus mirus* (Osad.), and fragments of trilobites, brachiopods, and *Hyolitha* spp.;

(7) ribbed, lenticular-bedded carbonate shales showing equal proportion of clay to carbonate and enclosing some interbeds of black clayey limestones (26 m); the unit bears trilobites and archaeocyathids of the Kameshki Horizon of the Atdabanian Stafe: *Bulaiaspis taseevica* Rep. (Tarnovskii, 1979), *Nochoroicyathus* sp., *Arturocyathus* cf. *borisovi* Roz., *Nalivkinicyathus cyroflexus* (Bojar. et Osad.), and *Erismacoscinus* sp;

(8) alternating sandstones, siltstones, and dark-gray limestones (up to 50 m); limestones yielded archaeocyathids of the Sanashtykgol Horizon of the Botomian Stage: *Irinaecyathus ratus* (Zhur.), *Tegerocyathus* sp., *Capsulocyathus* sp., *Coscinocyathus simplex* Vol., *Clathricoscinus sanaschtykgolensis* Borod. et Osad.,

Anthomorpha rackovskii (Vol.), *Shivegocyathus plenus* Fonin, *Archaeocyathus* sp. and some *Hyolitha* forms.

Further along the river valley, there is an exposure of light-red massive reef limestones 300–350 m thick (Fig. 2). This extensive biohermal massif constitutes a positive structure with bed dipping from the medial axis toward the south and north. The further bed sequence is as follows;

(9) light-gray platy clastic limestones (60 m) with archaeocyathids of the Kiya Horizon: *Nochoroicyathus khemtschikensis* (Vol.), *Plicocyathus* sp., *Gordonicyathus gerassimovensis* (Krasn.), *Archaeolynthus unimurus* (Vol.); this fauna coexists with cribricyatheans *Szecyathus cylindricus* (Vol.) and renalcids of genera *Epiphyton* and *Tubomorphophyton*;

(10) light-pink massive bun-shaped limestone bodies as large as 0.5 × 1.2 m (10 m); the rocks bear archaeocyathids of the Kiya and Kameshki horizons: *Plicocyathus mirus* (Osad.), *Tumulocyathus* sp., *Capsulocyathus* sp., *Nalivkinicyathus cyroflexus* (Bojar. et Osad.), *Arturocyathus borisovi* Roz., *Nochoroicyathus khemtschikensis* (Vol.), *Erismacoscinus* sp., *Loculicyathus membranivestites* Vol., and *Archaeolynthus* sp.;

(11) light biohermal limestones (25 m) with archaeocyathids of the Kameshki Horizon: *Gordonicyathus howelli* (Vol.), *Clathricoscinus sanaschtykgolensis* Borod. et Osad., *Retecoscinus retetabulae* (Vol.), *Plicocyathus mirus* (Osad.), *Torosocyathus inclinatus* Osad., *Nochoroicyathus* sp., *Capsulocyathus* sp., *Dictyocyathus* sp., *Tumulocyathus* sp., *Dokidocyathus* sp.; the assemblage coexists with renalcids of genera *Epiphyton, Tubomorphophyton, Renalcis, Razumovskia,* and *Girvanella*, and with small-shelled fauna of *Hyolithellus* sp., *H. vladimirovae* Miss., *H. tenuis* Miss., *Torellella* sp., *T. biconvexa* Miss., *T. lentiformis* (Sys.), *Rhombocorniculum cancellatum* (Cobb.), *R.* cf. *walliseri* Mamb., *Mongolitubulus squamifer* MIss., *Tommotia plana* (Miss.), and *Lapworthella nigra* Cobb.;

(12) light biohermal limestones (15 m); fossils are represented by archaeocyathids of the Sanashtykgol Horizon: *Kijacyathus chomentovskii* Z. Zhur., *Coscinocyathus* sp., *Anthomorpha* sp., *Sajanocyathus ussovi* Vol., *Capsulocyathus irregularis* (Zhur.), *Sibirecyathus dissepimentalis* (Vol.), *Thalamocyathus* sp., *Aptocyathus* sp., *Anthomorpha ? sisovae* (Vol.); by renalcids of genera *Epiphyton, Tubomorphophyton, Renalcis, Proaulopora, Razumovskia*, and *Botominella*; and by small-shelled fauna of *Hyolithellus tenuis* Miss., *H. vladimirovae* Miss., *Torellella* sp., *T. lentiformis* (Sys.), *T. biconvexa* Miss., *Mongolitubulus squamifer* Miss., *Rhombocorniculum* sp., *R. cancellatum* (Cobb.), *R. walliseri* Mamb., *R.* sp., *Tommotia plana* (Miss.), and *Lapworthella nigra* Cobb.;

(13) dark clayey limestones and carbonate shales with abundant trilobite fragments (10–12 m); archaeocyathids of the Sanashtykgol Horizon from the rocks are *Archaeolynthus* sp., *Antomorpha rackovskii* (Vol.), *Clathricoscinus sanaschtykgolensis* Borod. et Osad., *Shiveligocyathus plaenus* Fonin, *Coscinocyathus simplex* Vol., *Irinaecyathus ratus* (Vol.), and *Archaeocyathus* sp.;

(13a) the northern slope of the Balakhtison reef: pale massive and flaggy cyanobacterial limestones (20 m) bearing archaeocyathids and trilobites of the Sanashtykgol Horizon, namely *Archaeolynthus crassimurus* (Vol.), *Dictyocyathus* sp., *Tumulocyathus* sp., *Loculicyathus tolli* Vol., *Clathricoscinus infirmus* (Vol.), *Tumuliolynthus* cf. *tubexternus* (Vol.), *Altaicyathus vologdini* (Yawor.) (identifications of I.T. Zhuravleva), *Bonnia* sp., and *Poliellina lermontovae* Polet. (identifications by Repina);

(14) greenish gray shales and gray clayey limestones (12 m) with small bioherms (biostromes) composed entirely of large archaeocyathids characterizing the Solontsy Subhorizon of the Obruchev Horizon of the Toyanian Stage: *Pycnoidocyathus erbiensis* (Zhur.), *Archaeocyathus* cf. *cumfundus* (Vol.), *Irinaecyathus ratus* (Vol.), and *Schiveligocyathus plenus* Fonin.

All deposits described above overlie variegated conglomerates, gravelstones, sandstones, and polymictic siltstones. The rhythmical alternation is characteristic of fine- and coarse-grained rock varieties. Pebbles contain archaeocyathids characteristic of all Lower Cambrian horizons. The rocks also enclose abundant pebbles of Ol'khovka granitoids 512–544 Ma old. We attributed these conglomerates to the Kotel Formation of presumable Middle Cambrian age.

It seems impossible at present to divide the Lower Cambrian deposits of the Balakhson River section into the Kolpa and Balakhtison formations. Analyzing lithological features and fossils from the section fragments, we inferred that the Balakhtison Formation of dark-gray rhythmically bedded carbonate shales enclosing small and big reefal buildups and the Torgashino Formation of the southern slope of the Torgashino Ridge in the west of Eastern Sayan were formed concurrently during the time span of the Kiya, Kameshki, and Sanashtykgol horizons.

According to the available records, the upper Chibizhek Formation underlying the Balakhtison deposits can be attributed to the Lower Cambrian and correlated with the Ungut Formation of the Mana trough (see below).

THE CHIBIZHEK STRATOTYPE AREA

The fourth area of distribution of the Lower Cambrian deposits is situated to the southwest of the Balakhtison area, on the right bank of the Chibizhek River head. In the north it bounds the Ol'khovka granitoid intrusion (Figs. 1, IV and 3a).

The Lower Cambrian Chibizhek Formation was established in 1957 by Dodin *et al.*, (1959), who considered it as younger in age than the Balakhtison Formation.

Three Bezymyannyi, Chibizhek, and Balakhtison (Kolpa) stratigraphic units have been distinguished long ago in this area. Their stratigraphic position, scope and interrelations are still disputable.

Of undoubted Lower Cambrian age are the dark-gray and black carbonate shales with dark limestone interbeds. We attribute them to the Balakhtison Formation, whereas Vorob'ev *et al.* (unpublished report of 1989) consider them as rocks of the Kolpa Formation. Early Cambrian archaeocyathids and renalcids were found in different outcrops of these deposits. According to drilling and geological mapping data, the Vendian–Cambrian beds of the Chibizhek Formation become younger from contact with the Besymyannyi Formation in the north toward the Balakhtison Formation in the south; their dipping angle is not more than 30°. In 1988, we dated the Chibizhek Formation more precisely and subdivided it into the lower and upper subformations.

Two sections of the Chibizhek Formations are described below.

The section along the line A–A (Fig. 3) from the north to the south consists of units (from the base upsection) described below.

The Bezymyannyi Formation:

(1) variegated terrigenous-tufogenous deposits (over 500 m).

The Chibizhek Formation, lower subformation:

(2) light-gray to white and dark-gray dolomites and dolomitic limestones with massive interbeds of limestones and cherts (500 m); according to Tarnovskii (1979), the beds yielded phytoderivates *Algotactis kabirsaensis* Posp., *A.* aff. *sinzasiensis* Posp., *Anabarostroma* sp., *Minjarostroma* sp., *Nodulosistroma* sp., *Sinsasephyton* sp., *Spongistroma* sp., *Bogojulia* sp., *B.* cf. *abnirmis* Posp., *Makarakistroma* sp., *Olekmia* sp., *Sarmaella* sp. and microphytoliths *Ambigolamellatus* sp., *Osagia* aff. *senta* Z. Zhur., *O.* cf. *tolerabila* Yaksch., *V.* cf. *compactus* Yaksch., *V. textus* Klunger, *V.* cf. *simplaris* Yaksch., and *Volvatella* sp., all characteristic of the West Siberian Horizon of the Upper Riphean.

The Chibizhek Formation, upper subformation:

(3) calcareous conglomerates and clastic limestones (calcarenites) with pebbles of diverse dolomites and limestones (up to 30 m); dolomites contain microphytoliths characteristic of the Yudomian Stage of the Vendian, such as *Vesicularites* aff. *concretus* Z. Zhur., *V. concretus* Z. Zhur., *Vermiculites* sp., and *Nubecularites* sp.;

(4) dark-gray bedded to laminated limestones alternating with cloddish clayey limestones and enclosing rare "calcarenite" interbeds (85-90 m); the rocks yielded microphytoliths *Nubecularites catagraphus* Reitl., *N. punctatus* Reitl., and *Glebosites* sp., renalcids *Renalcis* cf. *polymorphus* (Maslov) Reitl. and *Girvanella* sp., and small-shelled fauna of *Torellella* sp. and *Hyolitellus* sp.;

(5) light-gray lenticularly bedded limestones alternating with microphytolithic and more massive. cloddish limestones (100 m); microphytoliths *Nubecularites catagraphus* Reitl., and *N. punctatus* Reitl., and *Hieroglyphites mirabilis* Reitl. coexist with renalcids *Batinevia* sp. within the unit.

The Balakhtison Formation:

(6) dark-gray carbonate-clayey shales and siltstones with black limestone interbeds (up to 50 m); The deposits contain rare archaeocyathids *Sibirecyathus* sp., and renalcids of genera *Botomaella, Girvanella,* and *Epiphyton.*

The section along the line B–B (Fig. 2):

The Chibizhek Formation, lower subformation:

(1) gray and light-gray dolomites with interbeds of cherts and brecciated dolomites (up to 300 m); microphytoliths *Nubecularites antis* Z. Zhur., *Ambigolamellatus horridus* Z. Zhur., *Volvatella* aff. *zonalis* Nar., *Vasicularites artus* Yaksch., *Vermiculites* ex gr. *liris* Reitl., *V. tortuosus* Reitl., and *Oncostroma* sp. are typical of the unit.

The Chibizhek Formation, upper subformation:

(2) brecciated dolomitic limestones alternating with microphytolithic rock varieties (up to 200 m) and bearing microphytoliths *Osagia* ex gr. *grandis* Z. Zhur., *O. korbinica* Yaksch., *Ambigolamellatus horridus* Z. Zhur., *Volvatella* aff. *zonalis* Nar., and *Vermiculites angularis* Reitl.;

(3) alternation of cloddish and dolomitic limestones (50 m); fossils from the unit are renalcids of genera *Epipyton, Botomaella, Girvanella, Kordephyton,* and small-shelld fauna of *Torellella* sp., *T. lentiformis* (Sys.), *T. curva* Miss., *Coleoloides* sp., *Anabarites* sp., and *Glauderia* sp.

The Balakhtison Formation:

(4) gray rhythmically bedded calcareous sandstones alternating with greenish gray to black siltstones (200 m);

(5) gray clastic and conglomerate-like limestones with gritstone and sandstone interbeds (up to 10 m); archaeocyathids *Propriolynthus* sp., *Nochoroicyathus* sp., cribricyathids, and renalcids of genera *Epiphyton, Proaulopora,* and *Renalcis* represent fossils of the unit.

On the right bank of the Egor'evka River (the left tributory of the Dzhebi River), the Upper Chibizhek Subformation includes limestones containing the Early Cambrian renalcids *Razumovskia uralica* Vol., *Epiphyton inexpectatum* Korde, *E. frondosum* Korde, and *Renalcis pectunculus* Korde identified by Luchinina in 1989.

CORRELATION OF THE BASAL BEDS OF THE LOWER CAMBRIAN

The Lower Cambrian sequences of four areas can be easily correlated on the basis of archaeocyathid and trilobite assemblages (Fig. 2). Oldest archaeocyathids and associated small-shelled fauna are referred to the

Natal'evka Horizon of the lowermost Atdabanian Stage in the Altai–Sayan foldbelt. These archaeocyathids of the Tayaty or Nizhnyaya Bazaikha assemblage (Musatov *et al.*, 1961) considerably differ from the Bazaikha archaeocyathid assemblage of the stratotype section. The former assemblage predominantly consists of simple morphological forms of *Archaeolynthus, Nochoroicyathus, Rotundocyathus, Dictyocyathus*, and other genera. Beginning from the first meters of the section, the archaeocyathids are associated with small-shelled fauna of *Hyolithellus* sp., *H. tenuis* Miss., *H.* cf. *isiticus* Miss., *H.* ex gr. *isiticus* Miss., *Torellella* sp., *Sunnaginia* sp., *S.* cf. *imbricata* MIss., *Coleolella* sp., *Coleolus* sp., *C.* cf. *trigonus* Sys., and also with chancelloriids and brachiopods. In addition, the analogues of the Ungut Formation in the right bank of the Yenisei River yielded problematic skeletal fossils *Cloudina* sp. (Fig. 4), which were earlier recovered together with *Renalcis* from the Zabit Formation of the Bokson Group of the Eastern Sayan (Kheraskova and Samygin, 1992).

Two lower beds of the Sretenka Formation (240 m), which are represented by gray and dark-gray bedded limestones with rare interbeds of calcareous and polymictic sandstones, are well correlated with the Kaltat Member of the Torgashino section on the base of lithological features. However, the Kaltat deposits contain unidentifiable fragments of archaeocyathids. The overlying red Bazaikha Member bears the younger Kiya archaeocyathid assemblage, and the Kaltat beds enclose the small-shelled funa dominated by Hyolithelminthus forms. The occurrence of archaeocyathid fragments may indicate the Atdabanian (Natal'evka) age of the Kaltat Member. In any case, the beds with these fossils are not older than the Tommotian, although they contain *Siphogonuchites* sp., which are commonly considered to be typical of the Nemakit-Daldynian Age.

The Kaltat and Ungut formations are closely related with terrigenous graywackes of the Anastas'ino Formation of the Mana trough and constitute the upper part of a single transgressive megacyclothem (Fig. 1, I, V). Most likely, the Anastas'ino Formation is of the Vendian age, because it unconformably overlies dolomites of the Angaloi Formation containing microphytoliths and acritarch assemblage of the Upper Riphean (the West Siberian Horizon).

In Shipitsyn's opinion, the association of renalcids and microphytoliths from the Upper Chibizhek Subformation is correlatable with that of the Ungut Formation of the Mana trough. The concurrent age of these deposits is also supported by occurrence of *Torellella* sp., *T. lentiformis* (Sys.), *T. curva* Miss., *Hyolithellus* sp., *Coleoloides* sp., *Anabarites* sp., and *Glauderia* sp. in the upper Chibizhek Subformation.

The Ungut, Kaltat, and upper Chibizhek deposits rest on the beds referred to the Vendian. So the Tommotian Stage of the northwestern Eastern Sayan may be represented by the lower Kaltat and lower Ungut formations, and even by upper beds of the Anastas'ino Formation. Analogues of the Tommotian deposits appear to be lacking to the south and southwest of the Torgashino Ridge and the Mana trough, as it is evident from the transgressive type of Early Cambrian sedimentation in the region, from where the terrigenous-carbonate deposits similar to the Anastas'ino Formation are missing. A much younger age of the carbonate shales of the Balakhtison Formation and other sections near the Sretenski Reef is indicated by enclosed fossils.

Fig. 4. *Cloudina* sp. from the Kaltat Formation of the Torgashino Ridge, (×10).

Dolomites with the Vendian microphytoliths, which underlie the Upper Chibizhek Subformation, should be correlated to the Ovsyanka and Angaloi formations beneath the Anastas'ino Formation.

CONCLUSION

(1) The coexistence of archaeocyathids characteristic of the Natal'evka Horizon and small-shelled fauna in the lower beds of the Sretenka Formation allows us to specify the age of the Lower Cambrian deposits exposed in sections correlative with the Sretenka section. According to our correlation scheme, the Kaltat Member of the Torgashino section, as well as the upper beds of the Ungut Formations and upper Chibizhek Subformation, are attributable to the Natal'evka Horizon of the Atdabanian Stage.

(2) The reliable correlation of the Lower Cambrian deposits on the basis of archaeocyathids and trilobites is possible beginning from the Kiya Horizon, as the sea transgression spread over the region at that time.

(3) The expansion of the sea basin continued till the mid-Botomian, when the cyanobacterial–archaeocyathid reefs were prosperous. In the interval of Kiya, Kameshki, and lower Sanashtykgolian horizons, there are no large hiatuses corresponding to stratigraphic zones or subzones.

(4) In the second half of the Sanashtykgolian time, the sea basin reduced, sedimentation turned out to be terrigenous, and some deposits were washed out. In

massive reefs, erosion caused the missing of beds corresponding to zones and subzones.

(5) The Early Cambrian transgression peaked most likely in the Tommotian time. The occurrence of typical Tommotian forms of small-shelled problematics in the younger beds of the Eastern Sayan suggests a great intensity of erosion and redeposition in the Early Cambrian.

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The Bed-by-Bed Standard of the Silurian for East Siberia: Fauna and Flora

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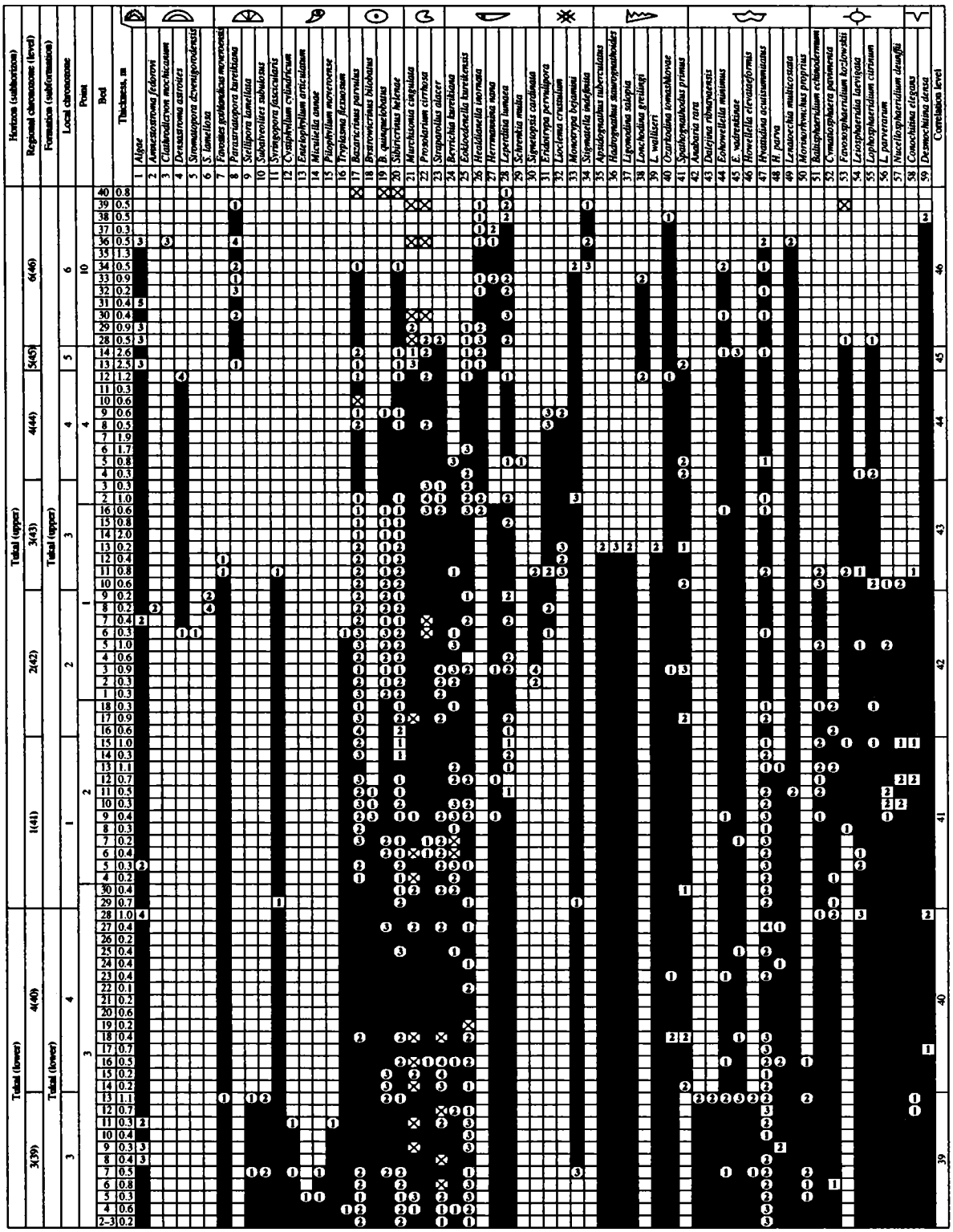
Abstract—The bed-by-bed standard of the Silurian for East Siberia is used to correlate stratigraphic sequences all over the Siberian platform and Taimyr. Fossil faunas and floras of the standard are very representative, because the original type sections of the standard consist of deposits accumulated in shallow shelf settings favorable for biota evolution. Consecutive beds of the standard are characterized by distribution parameters (abundance rates and size ranges) of the following fossil groups: algae, stromatoporoids, tabulates, heliolitids, rugose corals, crinoids, gastropods, ostracodes, tentaculitids, bryozoans, conodonts, scolecodonts, bivalves, brachiopods, trilobites, cephalopods, acritarchs, chitinozoans, and graptolites. The biocoenotic analysis is based on data characterizing the taxonomy and relative abundance of fossils in oryctocoenoses, and on the structure of the latter.

Key words: *Siberia, fauna, flora, standard, oryctocoenosis.*

The bed-by-bed standard of the Silurian for East Siberia (Siberian platform and Taimyr) is a composite section incorporating stratotypes of regional chronozones in their stratigraphic succession. Particular combinations of chronozones represent stratotypes of East Siberian subhorizons and horizons (Tesakov *et al.*, 1996a, 1996b, 1998a, 1998b). Data on stratotype localities in the Moierokan (Moiero River basin), Turukhansk (Kureika River basin), and Norilsk regions (the eastern coast of the Pyasino Lake) have been published earlier together with large-scale maps and a detailed lithological description (Tesakov *et al.*, 1998a, Figs. 1–3). The stratotypes in question are most representative in a paleontological aspect, because they correspond to sedimentary sections accumulated in the open shelf settings. The only exception is the stratotype of the Postnichnyi Horizon (Ludfordian and Přidoli) that is composed of lagoonal sedimentary facies almost barren of organic remains. In order to characterize this stratigraphic level, we used the parastratotype of the Nizhnyaya Taimyra River, where beds are rich in marine fauna (Tesakov *et al.*, 1995, Exposure 218).

In our earlier works, floras and faunas of the bed-by-bed standard have been described only for individual sections of the Moiero (Tesakov *et al.*, 1985, 1986) and Kureika river basins (Tesakov *et al.*, 1980) without necessary coordination with regional chronozones, and these descriptions were of a low significance for correlation procedure. The following research team contributed much to investigation of individual groups of fossils: V.G. Khromykh (stromatoporoids), Yu.I. Tesakov (tabulates and heliolitids), A.B. Ivanovskii and Yu.Ya. Latypov (rugose corals), G.A. Stukalina (crinoids), N.I. Kurushin and A.P. Gubanov (gastropods), L.S. Bazarova (ostracodes), A.Ya. Berger (tentaculitids), K.N. Volkova (bryozoans), T.V. Mashkova and T.A. Moskalenko (conodonts), N.M. Zaslavskaya (scolecodonts and chitinozoans), T.V. Lopushinskaya (brachiopods), E.A. Elkin (trilobites), E.I. Myagkova and O.K. Bogolepova (cephalopods), L.I. Sheshegova (acritarchs), A.M. Obut and N.V. Sennikov (graptolites). The presented data on abundance rates and size ranges of specimens are corrected in accordance with taphonomic studies of natural outcrops.

The bed-by-bed distribution of organic remains is shown in Fig. 1, where, in accordance with our subdivision scheme (Tesakov *et al.*, 1996, Table 2; 1997, Table 3), we plotted stratigraphic ranges of the following subhorizons: lower Moierokan (lower–middle Rhuddanian), upper Moierokan (upper Rhuddanian–lower Aeronian), lower Khaastyr (basal middle Aeronian), upper Khaastyr (terminal middle–upper Aeronian), lower Agidy (lower–basal middle Telychian), upper Agidy (terminal middle–upper Telychian), lower Khakoma (Sheinwoodian), upper Khakoma (Homerian), lower Tukal (basal lower Gorstian), and upper Tukal (terminal lower–upper Gorstian). In the second column, the first figures denote numbers of regional chronozones and figures in brackets are the ordinal numbers of correlation levels corresponding to regional biostratigraphic ranges (biophases). The abundance rate of specimens is defined by using the five-mark



Stratigraphic distribution of fossils in the bed-by-bed standard of the Silurian for East Siberia: (1) graptolites; (2) chitinozoans; (3) acritarchs; (4) cephalopods; (5) trilobites; (6) brachiopods; (7) mud-eaters; (8) scolecodonts; (9) conodonts; (10) bryozoans; (11) tentaculitids; (12) ostracodes; (13) gastropods; (14) crinoids; (15) rugose corals; (16) tabulates and heliolitids; (17) stromatoporoids; (18) stromatolites; (19) abundance rate corresponding to single (1), rare (2), few (3), common (4), and abundant (5) specimens; (20) large, (21) medium-sized, and (22) small specimens; (23) species identified in the open nomenclature; (24) ordinal numbers of regional chronozone and (in brackets) of correlation level = regional biointerval of the Silurian in East Siberia.

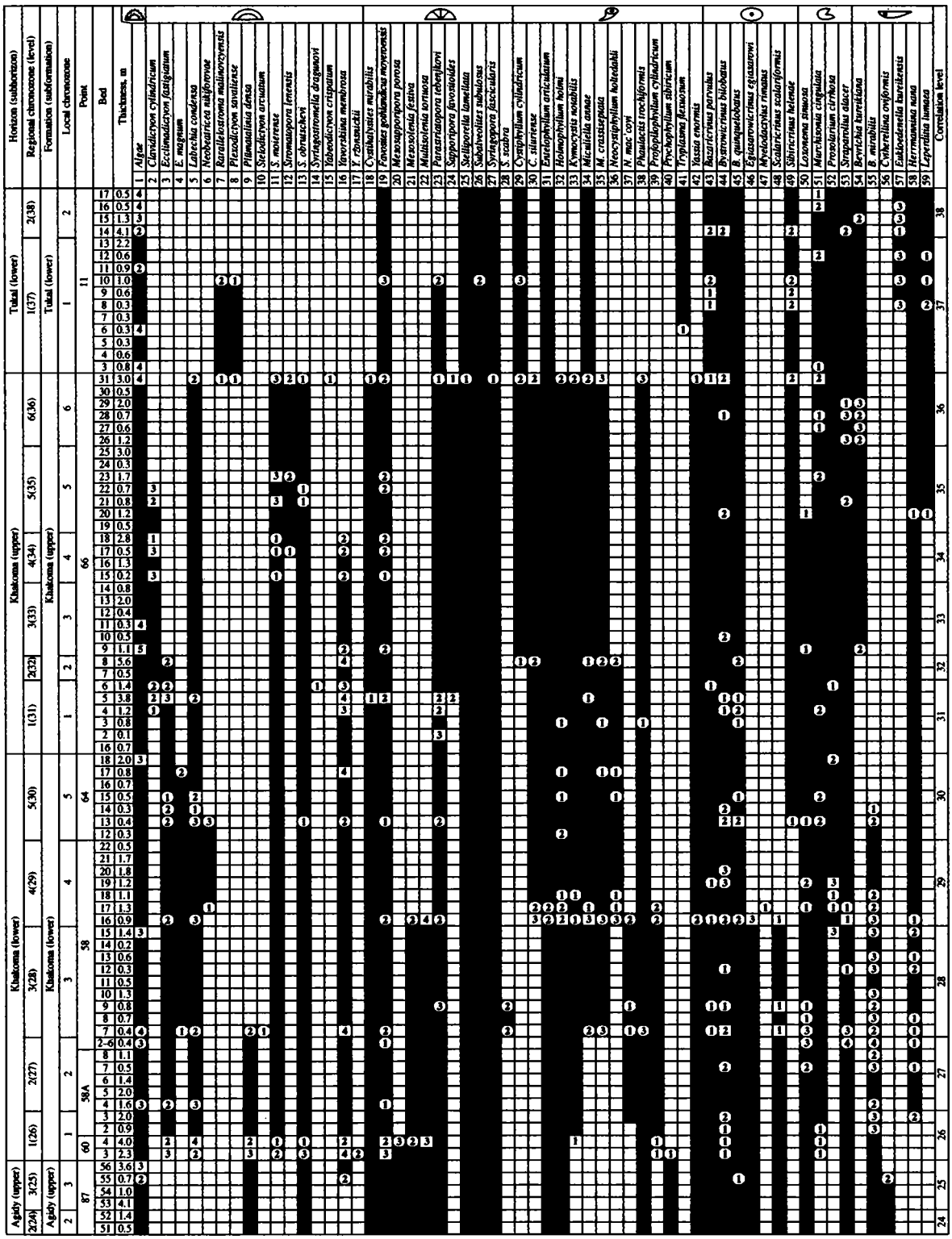
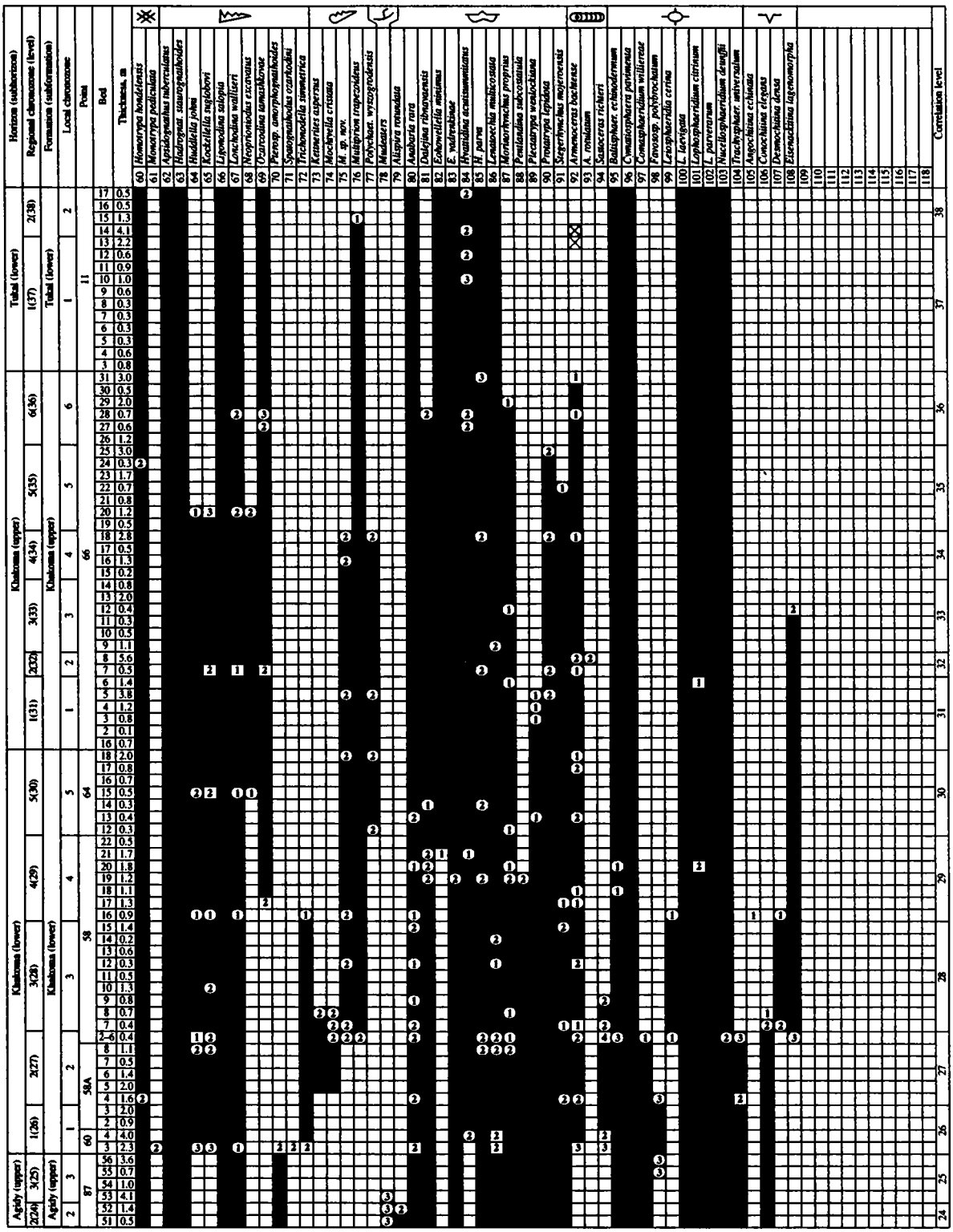


Figure. (Contd.)

scale: (1) single, (2) rare, (3) few, (4) common, (5) abundant. Open octahedrons, circles, and squares signify small, medium-sized and large specimens, respectively, and the black spaces mark intervals barren of fossils.

Beds shown in Fig. 1 are of the monofacies type in nature and correspond, in ecosystem terminology, to remains of parcellas or biogeocoenoses (Tesakov, 1978). Provided that biogeocoenosis is an elementary



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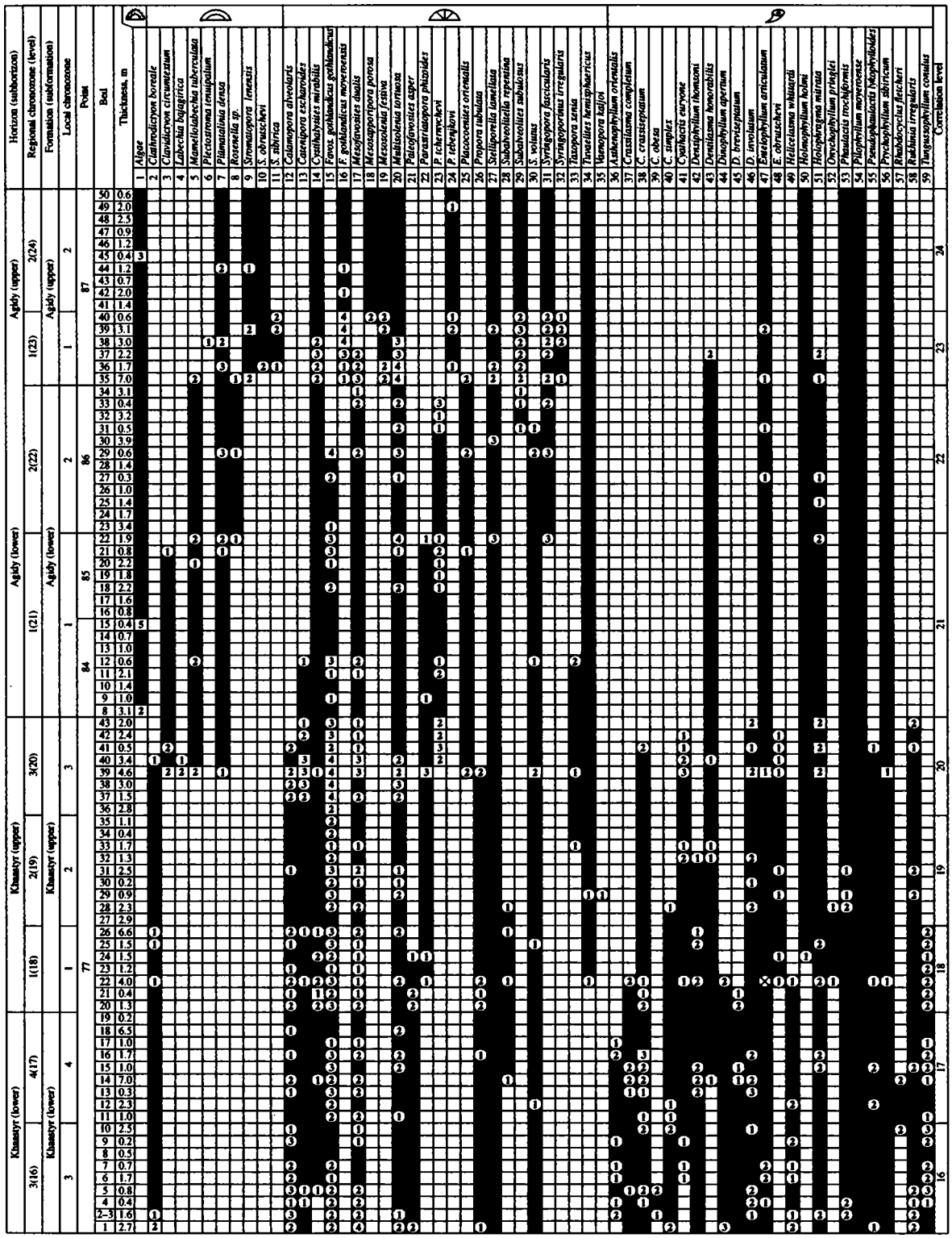
unit of ecosystems, we grouped fauna and flora together according to oryctocoenoses characterizing the bed groups within the range of remnant biogeocoenoses or taphonomic units of lithocoenoses.

The lower Moierokan Subhorizon is rich in graptolites, trilobites, brachiopods, and nautiloids; gastropods and ostracodes are less abundant here, and other faunal groups are scarce.



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(1) The *Alispira gracilis* oryctocoenosis (Bed 82/1) is enclosed into limestones and shales. Its structure is of the small-accumulation type. The principal oryctoparcella (limestones) shows a higher burial density as compared to the subordinate one (shales). Organic remains are small, and buried in the autochthonous mode.

(2) The *Geisonoceras kureikense* oryctocoenosis (Bed 82/2) is enclosed into limestones. It is homoge-

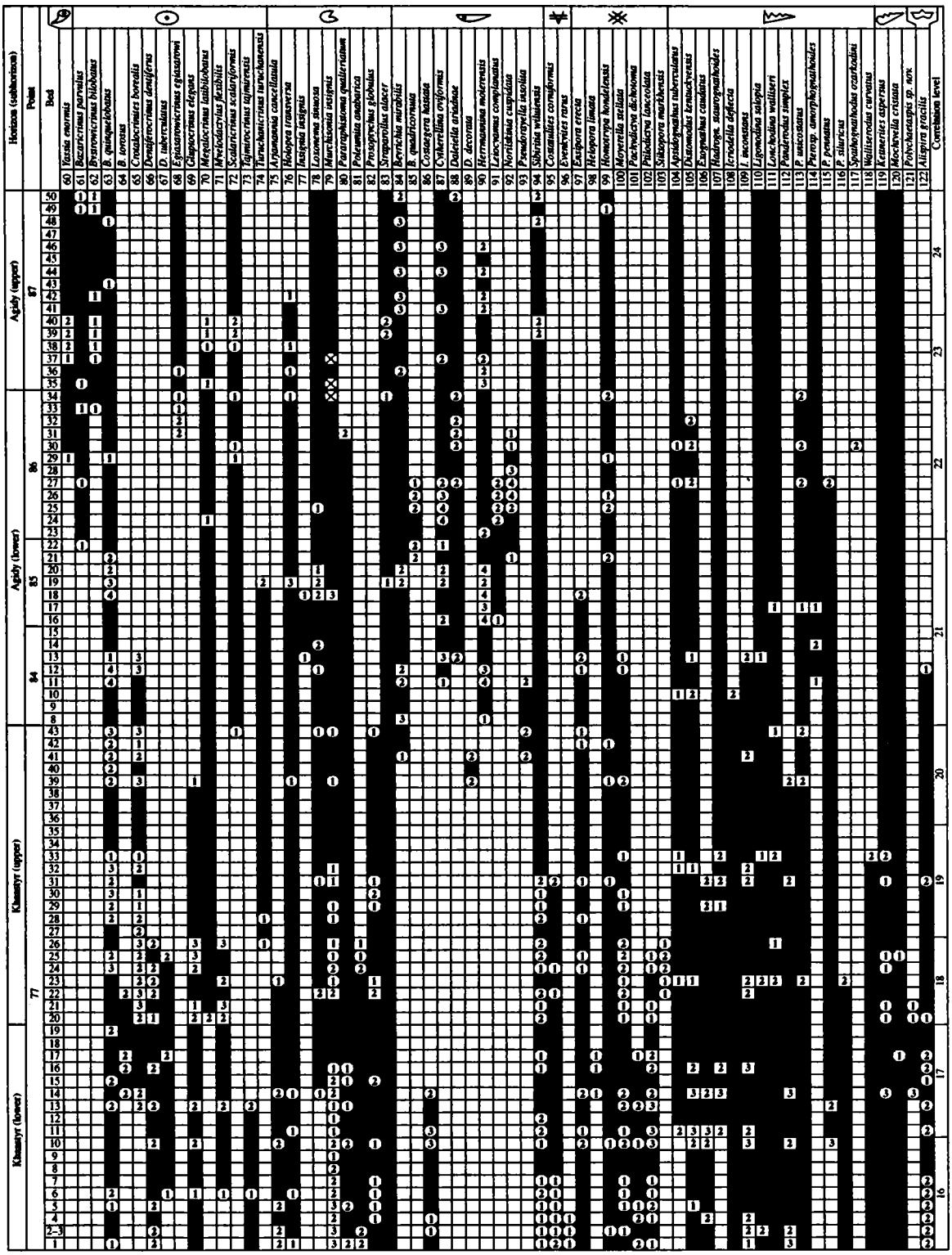
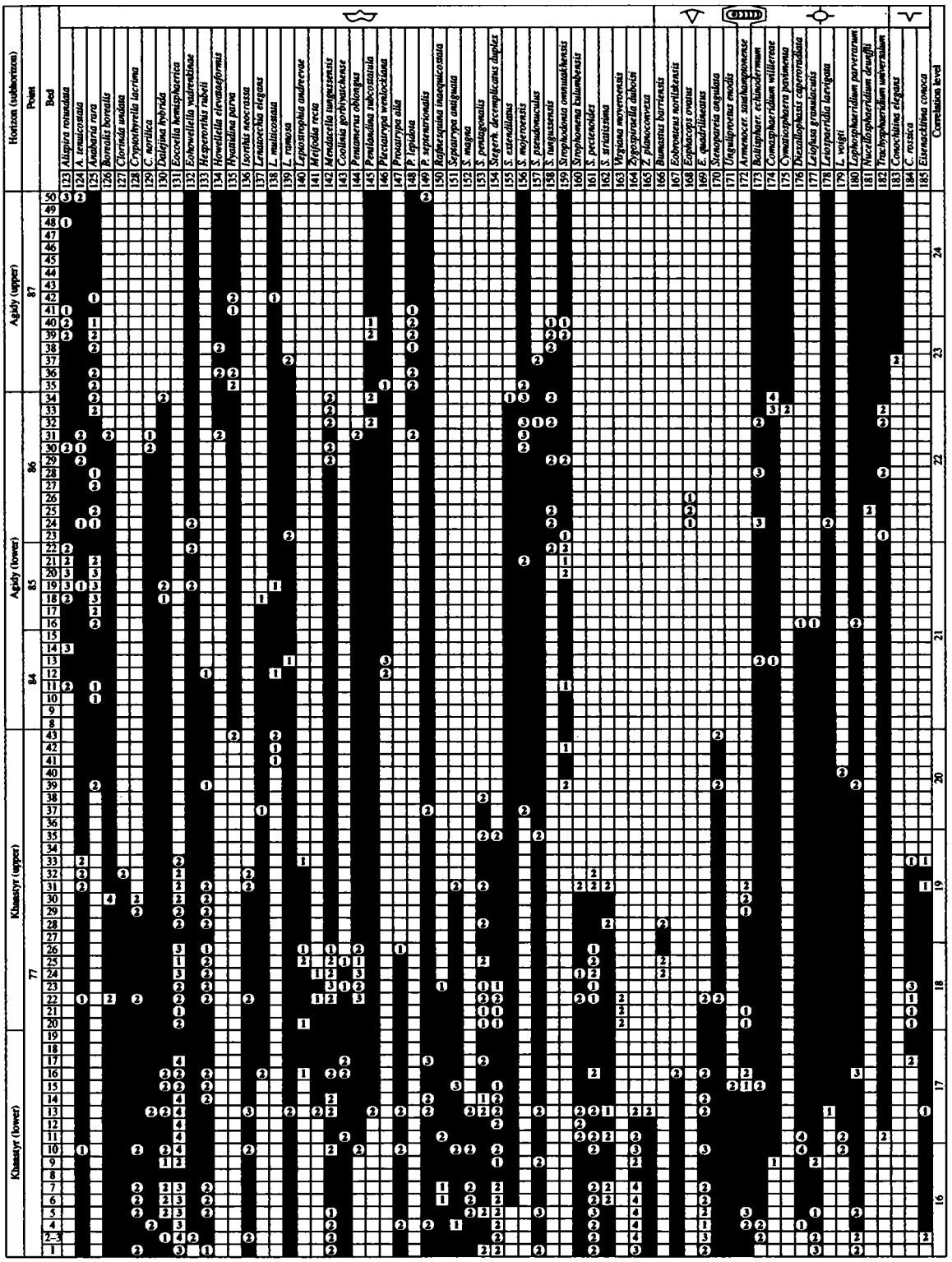


Figure. (Contd.).

neous in the structure. The burial density is very high. Organic remains are medium-sized and buried in the allochthonous, locally oriented mode.

(3) The *Glyptograptus tamariscus* oryctocoenosis (Bed 82/3) is enclosed into shales. The structure is banded. The burial density is moderate. Organic remains are small and buried mostly in the allochthonous mode either along bedding planes or chaotically.



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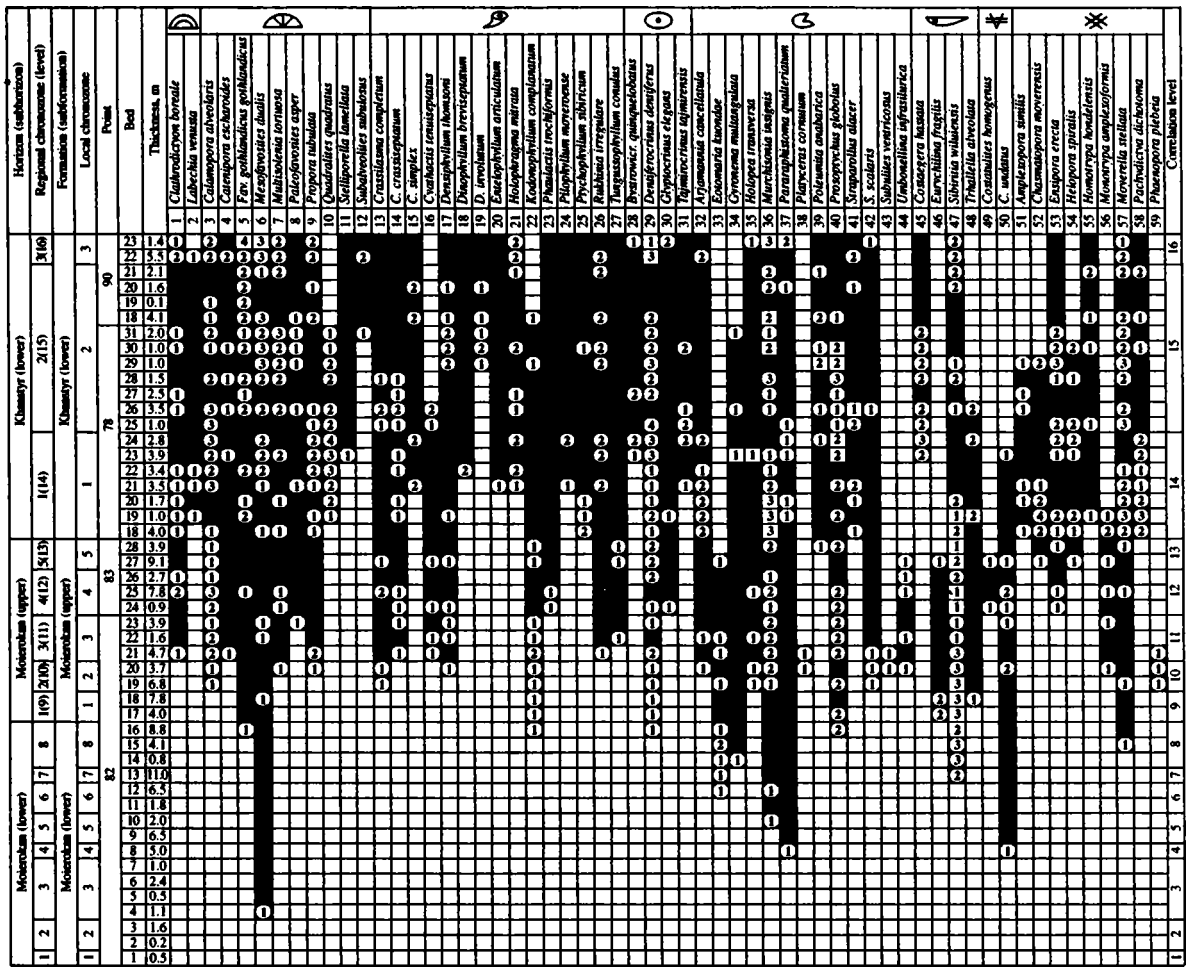
(4) The *Metabolograptus moyeroensis* oryctocoenosis (beds 82/4–82/7) is enclosed into shales and limestones. The structure is banded. In the main oryctoparcella (shales), the burial density is moderate, though more compact than in limestones. Organic remains are small and buried in the subautochthonous mode.

(5) The *Coronograptus cyphus* oryctocoenosis (beds 82/8–82/10) is enclosed into shales and lime-



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stones. The structure is banded. In the main oryctoparcella (shales), the burial density is moderate in contrast to the very low density in limestones. In shales, burials are confined to bedding planes. Organic remains are small and buried in a subautochthonous mode.

(6) The *Paraclimacograptus innotatus* oryctocoenosis (beds 82/11–82/12) is enclosed into limestones and shales. The structure is homogeneous. In both oryctoparcellas (limestones and shales), the burial density is very low. Organic remains are small and buried in a subautochthonous mode.

(7) The *Acernaspis superciliexcelsis* oryctocoenosis (beds 82/13–82/15) is enclosed into limestones and rare marl interbeds. The structure is homogeneous. The burial density is moderate. Organic remains are small and buried in the autochthonous mode.

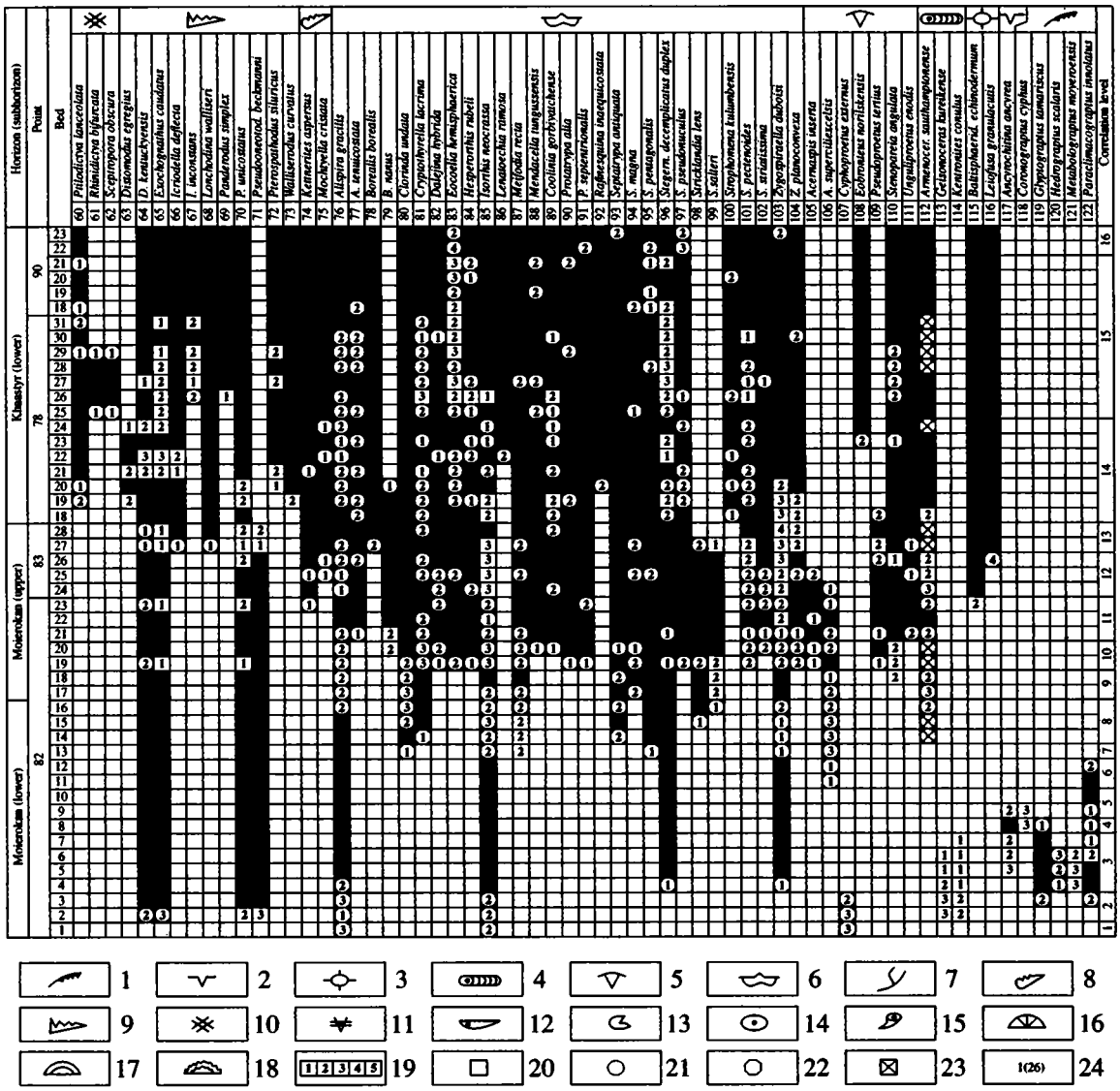
(8) The *Clorinda undata* oryctocoenosis (Bed 82/16) is enclosed into limestones. The structure is homogeneous. The burial density is low. Organic remains are small and buried in the autochthonous mode.

The upper Moierokan Subhorizon is characterized by widespread brachiopods and ostracodes; tabulates, rugose corals, crinoids, gastropods, trilobites, and cephalopods are less frequent, and other fossils occur as single specimens.

(1) The *Sibiritia wiluiensis* oryctocoenosis (beds 82/17–82/18) is enclosed into limestones and rare marl interbeds. The structure is homogeneous. The burial density is moderate. Organic remains are small, locally medium-sized, and buried in the autochthonous or subautochthonous mode.

(2) The *Septatrypa antiquata* oryctocoenosis (beds 82/19–82/20) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is moderate. Organic remains are small or medium-sized and buried in the autochthonous mode.

(3) The *Zygospiraella duboisi* oryctocoenosis (beds 82/21–82/23) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is moderate. Organic remains are medium-sized and generally buried in the autochthonous mode.



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(4) The *Isorthis neocrassa* oryctocoenosis (beds 83/24–83/26) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is high. Organic remains are small or medium-sized and buried in the autochthonous mode.

(5) The *Zygospiraella duboisi* oryctocoenosis (beds 83/27–83/28) is enclosed into cloddish and detrital limestones. The structure is of the small-cluster type. The burial density is moderate in the main oryctoparcella (cloddish limestones). Subordinate oryctoparcellas (lentils of detrital limestone) enclose the shell and detritus accumulations. Organic remains are small or medium-sized and generally buried in the autochthonous mode.

The lower Khaastyr Subhorizon yields abundant tabulates, rugose corals, crinoids, gastropods, bryozoans, and brachiopods; stromatoporoids, ostracodes, trilobites, and cephalopods are second in abundance, whereas other fossils occur as solitary specimens. The burial density is high. The size range of organic remains is moderate.

(1) The *Quadralites quadratus* oryctocoenosis (beds 78/18–78/24) is enclosed into cloddish, platty, and detrital limestones. The structure is of the small-cluster type. The burial density of autochthonous fossil accumulations is high in the basic oryctoparcella (cloddish to platty limestones). Subordinate oryctoparcellas (lentils of detrital limestone) are rich in accumulations of detritus, small shells, and taeniate bryozoans.

(2) The *Stegerhynchus decemplicatus duplex* oryctocoenosis (beds 78/25–78/28) is enclosed into cloddish limestones, marls, and lentils of detrital limestones. The structure is of the small-cluster type. The main oryctoparcella (cloddish limestones and marls)

shows the high burial density of medium-sized autochthonous organic remains. Subordinate oryctoparcellas (lentils of detrital limestone) enclose coquina and bryozoan accumulations.

(3) The *Mesofavosites dualis* oryctocoenosis (beds 78/29–78/31) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is high. Organic remains are small or medium-sized and buried in the autochthonous or subautochthonous mode.

(4) The *Eocoelia hemisphaerica* oryctocoenosis (beds 90/18–90/22) is enclosed into marls and lentils of limestones. The structure is of the small-cluster type. The main oryctoparcella (marls) shows the moderate burial density of small to medium-sized organic remains. Subordinate oryctoparcellas (lentils of limestones) enclose detritus and accumulations of small shells. Burials are of the autochthonous and subautochthonous types.

(5) The *Calamopora alveolaris* oryctocoenosis (beds 90/23, 77/1–77/10) is enclosed into cloddish limestones and lentils of detrital limestones. The structure is of the small-cluster type. The main oryctoparcella (cloddish limestones) shows the high burial density of small to medium-sized organic remains. Subordinate oryctoparcellas (lentils of detrital limestone) enclose accumulations of small valves. Burials are of the autochthonous and subautochthonous type.

(6) The *Eocoelia hemisphaerica* oryctocoenosis (beds 77/11–77/14) is enclosed into marls and lentils of detrital limestones. The structure is of the small-cluster type. The main oryctoparcella (marls) exhibits the moderate burial density of small organic remains. Subordinate oryctoparcellas (lentils of detrital limestone) enclose detritus of small shells. Burials are of the autochthonous type in general.

(7) The *Favosites gothlandicus gothlandicus* oryctocoenosis (beds 77/15–77/16) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is high. Organic remains, small to moderate in size, occur in autochthonous burials.

(8) The *Eocoelia hemisphaerica* oryctocoenosis (beds 77/17–77/19) is enclosed into marls and lentils of detrital limestones. The structure is of the small-cluster type. The main oryctoparcella (marls) demonstrates frequent autochthonous burials of small organic remains. Subordinate oryctoparcellas (detrital limestones) enclose shell detritus and coquinas. Burials are of the autochthonous and subautochthonous types.

The upper Khaastyr Subhorizon is rich in brachiopods and crinoids; stromatoporoids, rugose corals, gastropods, ostracodes, bryozoan, and trilobites are less abundant, whereas other fossils are scarce.

(1) The *Pentamerus oblongus* oryctocoenosis (beds 77/20–77/25) is enclosed into marls intercalated with lentils of detrital limestones. The structure is of the small-cluster type. The main oryctoparcella (marls) shows the moderate density of autochthonous burials of small to medium-sized fossils. Subordinate oryctoparcellas (lentils of detrital limestone) enclose abundant small shells occurring in the subautochthonous position.

(2) The *Eocoelia hemisphaerica* oryctocoenosis (Bed 77/26) is enclosed into marls, cloddish limestones and lentils of detrital limestones. The structure is of the small- to medium-cluster type. In the main oryctoparcella (marls), the burial density of small autochthonous organic remains is moderate. Subordinate oryctoparcellas (cloddish limestones) are rich in autochthonous fossils of medium size range, whereas lentils of detrital limestone enclose small dispersed shells.

(3) The *Favosites gothlandicus gothlandicus* oryctocoenosis (beds 77/28–77/29) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is low. Organic remains of the moderate size occur in autochthonous burials.

(4) The *Eocoelia hemisphaerica* oryctocoenosis (beds 77/30–77/35) is enclosed into marls, cloddish limestones and lentils of detrital limestones. The structure is of the small- to medium-cluster type. In the main oryctoparcella (marls), the burial density of small autochthonous organic remains is moderate. Subordinate oryctoparcellas (lentils of detrital and cloddish limestones) are rich in small allochthonous fossils.

(5) The *Favosites gothlandicus gothlandicus* oryctocoenosis (beds 77/36–77/43) is enclosed into platty, cloddish, and laminated limestones. The structure is banded. The main oryctoparcella (marls) shows the low density of autochthonous burials of medium-sized corals. Subordinate oryctoparcellas (cloddish limestones) enclose frequent clusters of large autochthonous corals. In laminated limestones, the burial density is very low.

The lower Agidy Subhorizon is marked by widespread ostracodes, brachiopods, and tabulates, which associate with less frequent stromatoporoids, rugose corals, crinoids, gastropods, bryozoans, trilobites, and cephalopods; other fossils are sparse.

(1) The *Herrmannina moierensis* oryctocoenosis (beds 84/8–84/12) is enclosed into laminated and platty limestones. The structure is banded. The main oryctoparcella (laminated limestones) shows the extremely low density of allochthonous burials of small organic remains. Subordinate oryctoparcella (platty limestones) encloses quite frequent autochthonous fossils medium in size.

(2) The *Cytherellina oviformis* oryctocoenosis (beds 84/13–84/15, 85/16–85/19a) is enclosed into marls and platty limestones. The structure is of the small-cluster type. The basic oryctoparcella (marls) shows the moderate density of autochthonous burials of small organic remains. Subordinate oryctoparcellas (platty limestones) enclose subautochthonous clusters of small shells and algal buildups.

(3) The *Bystrowicrinus quinquelobatus* oryctocoenosis (beds 85/19b–85/19d) is enclosed into cloddish and platty limestones. The structure is of the cluster type. The basic oryctoparcella (cloddish limestones)

are characterized by the moderate density of autochthonous burials of small organic remains. The subordinate oryctoparcella (platty limestones) bears no abundant but very diverse fossils.

(4) The *Herrmannina moierensis* oryctocoenosis (Bed 85/20) is enclosed into platty limestones. The structure is homogeneous. The burial density is high. Autochthonous fossils are large to medium in size.

(5) The *Multisolenia tortuosa* oryctocoenosis (beds 85/21–85/22) is enclosed into cloddish limestones. The structure is homogeneous. The burial density is moderate. Autochthonous organic remains are medium in size.

(6) The *Cytherellina oviformis* oryctocoenosis (beds 86/23–86/29) is enclosed into marls and limestone lenses. The structure is of the small-cluster type. The basic oryctoparcella (marls) shows the moderate density of autochthonous burials of small organic remains. Subordinate oryctoparcellas (limestone lenses) are rich in allochthonous small fossils.

(7) The *Mendacella tungussensis* oryctocoenosis (beds 86/30–86/34) is enclosed into marls and lentils of detrital limestones. The structure is of the medium- to small-cluster type. The major oryctoparcella (marls) exhibits the low density of autochthonous burials of small organic remains. Subordinate oryctoparcellas enclose concentrated autochthonous burials and less frequent allochthonous clusters of fossils.

The upper Agidy Subhorizon is characterized by widespread ostracodes, brachiopods, and tabulates; stromatoporoids, crinoids, rugose corals, and gastropods are less abundant; other types of fossils occur as single specimens.

(1) The *Multisolenia tortuosa* oryctocoenosis (beds 87/35–87/36) is enclosed into cloddish limestones and bioherms. The structure is of the dispersed-cluster type. The major oryctoparcella (cloddish limestones) shows the high density of autochthonous burials of medium-sized organic remains. Subordinate oryctoparcellas (bioherms) are very rich in autochthonous tabulates and stromatoporoids.

(2) The *Favosites gothlandicus gothlandicus* oryctocoenosis (Bed 87/37) is enclosed into marl beds with limestone segregations. The structure is homogeneous, with moderately frequent burials of the medium-sized autochthonous organic remains.

(3) The *Favosites gothlandicus moyeroensis* oryctocoenosis (beds 87/38–87/40) is enclosed into cloddish limestone beds. The structure is homogeneous, with high density of burials. The medium-sized to large organic remains are autochthonous.

(4) The *Cytherellina oviformis* oryctocoenosis (Bed 87/41) is enclosed into shale and limestone beds. The structure is banded. Rare and very small autochthonous fossils occur in the main oryctoparcella (shales) only.

(5) The *Beyrichia mirabilis* oryctocoenosis (Bed 87/42) is enclosed into cloddish limestone beds. The structure is homogeneous. The burial density of small autochthonous organic remains is very low.

(6) The *Cytherellina oviformis* oryctocoenosis (beds 87/43–87/47) is enclosed into marls and cloddish limestones. The structure is banded. The low density of burials of small autochthonous organic remains is characteristic of the main oryctoparcella (marls). Subordinate oryctoparcellas (limestone interbeds) are more rich in burials.

(7) The *Beyrichia mirabilis* oryctocoenosis (Bed 87/48) is enclosed into dolomitic limestones. The structure is homogeneous. The burial density of small autochthonous organic remains is very low.

(8) The *Alispira rotundata* oryctocoenosis (beds 87/49–87/52) is enclosed into marls intercalated with limestone lentils. The structure is of the small-cluster type. In the main oryctoparcella (marls), the burial density of small autochthonous organic remains is very low, but they are more frequent, though also autochthonous, in subordinate oryctoparcellas (limestone lentils).

(9) The *Mudeaters* oryctocoenosis (Bed 87/53) is enclosed into foliated dolomitic limestones. The structure is homogeneous. The burial density of small and rare ostracode shells is very low.

(10) The *Cytherellina oviformis* oryctocoenosis (beds 87/54–56a) is enclosed into dolomitic limestone beds. The density of burials of small allochthonous organic remains is very low.

(11) The *Mudeaters* oryctocoenosis (Bed 87/56b) is enclosed into marls intercalated with rare limestone lentils. In burials of a rather high density, fucoids coexist with acritarchs only.

The lower Khakoma Subhorizon is characterized by abundant stromatoporoids, tabulates, rugose corals, and crinoids. Gastropods, ostracodes, brachiopods, and cephalopods are second in abundance, and other fossils are rare.

(1) The *Labechia condensa* oryctocoenosis (beds 60/3–60/4) from biohermal limestones is of the framework mosaic structure. The density of burial is very high. Autochthonous and allochthonous organic remains are large.

(2) The *Beyrichia mirabilis* oryctocoenosis (Bed 58A/2) is enclosed into cloddish limestones. The density of burials of small autochthonous fossils is low.

(3) The *Herrmannina nana* oryctocoenosis (Bed 58A/3) is enclosed into thin-bedded limestones. The structure is homogeneous. Autochthonous burials consist of small organic remains.

(4) The *Ecclimadictyon fastigiatum* oryctocoenosis (Bed 58A/4) is enclosed into cloddish and massive limestones. The structure is of the small-cluster type. The density of burials of medium-sized autochthonous fossils in the main parcella (cloddish limestones) is moderate. Subordinate oryctoparcellas (massive lime-

stones) generally bear the medium-sized stromatolites and stromatoporoids in frequent autochthonous burials.

(5) The *Favososphaeridium polybrochatum* oryctocoenosis (beds 58A/5–58A/6) is enclosed into dolomitic marls. The rocks bear only rare acritarchs.

(6) The *Beyrichia mirabilis* oryctocoenosis (beds 58A/7–58A/8, 58/2–58/14) is enclosed into marls intercalated with limestone lentils. The structure is banded. The low density of burials of small autochthonous fossils is characteristic of the main oryctoparcella (marls). Subordinate oryctoparcellas (limestone lentils) enclose rather frequent subautochthonous organic remains medium in size.

(7) The *Straparollus alacer* oryctocoenosis (Bed 58/15) from black and gray limestones is of the small-cluster structure. The density of burials of medium-sized autochthonous fossils is very high in the main oryctoparcella (black limestones). Subordinate oryctoparcella (gray limestones) is of a high burial density and consists of stromatolites.

(8) The *Beyrichia mirabilis* oryctocoenosis (beds 58/16–58/19) is enclosed into limestone and marls displaying wavy lamination. The structure is of the local small-accumulation type. The main oryctoparcella (limestones) shows the low density of subautochthonous burials of small fossils. Fossils of subordinate oryctoparcellas (marl interbeds) are represented by acritarchs only.

(9) The *Dalejina ribnayaensis* oryctocoenosis (Bed 58/20) is enclosed into marls with limestone lentils. The structure is of the small-accumulation type. The main oryctoparcella (marls) shows the low density of subautochthonous burials of small fossils. The same fossils occur in allochthonous position in subordinate oryctoparcellas (limestones).

(10) The *Dalejina ribnayaensis* oryctocoenosis from foliated limestones (beds 58/21–58/22) is homogeneous in structure. It consists of small brachiopod shells only.

(11) The *Labechia condensa* oryctocoenosis (beds 64/12–64/18) from platy limestones and biostromes is of the medium-cluster type. The main oryctoparcella (bedded limestones) shows the moderate density of autochthonous burials of small organic remains. Subordinate oryctoparcellas (biostromes) are rich in autochthonous stromatoporoids, stromatolites, and tabulates of medium size.

The upper Khakoma Subhorizon is found to be rich in stromatoporoids, tabulates, crinoids, and gastropods. These fossils coexist with less abundant cephalopods, bryozoans, brachiopods, and sporadically with singular specimens of other taxa.

(1) The *Yavorskiina membrosa* oryctocoenosis (beds 66/1b–66/8) is enclosed into platy limestones and bioherms. Its structure is of the large-cluster type. The main oryctoparcella (platy limestones) bears rare autochthonous organic remains of small and medium size. Subordinate oryctoparcellas (bioherms) enclose more frequent allochthonous and autochthonous burials of large stromatoporoids, tabulates, rugose corals, and crinoids.

(2) The *Bystrowicrinus bilobatus* oryctocoenosis (beds 66/9–66/12) is enclosed into massive and platy limestones. It is of the small-accumulation structure. The main oryctoparcella (massive limestones) shows the low density of allochthonous burials of small organic remains and abundant calyptras of stromatolites. Rare small-shelled fauna is concentrated in subordinate oryctoparcellas (platy limestones).

(3) The *Mudeaters* oryctocoenosis (beds 66/13–66/15) is enclosed into cloddish, platy, and biostromal limestones. Its structure is banded. Cloddish limestones yield rare and small brachiopod shells only. *Clavidictyon cylindricum* is dominant species in biostromes, where the burial density of medium-sized organic remains is high.

(4) The *Clavidictyon cylindricum* oryctocoenosis (beds 66/19–66/24) is enclosed into foliated and platy limestones and marls in a form of small accumulations and small clusters of fossils. The foliated limestones are almost barren of fossils and bear only rare crinoid columnals. Abundant autochthonous burials in marls and biostromes consist of medium-sized organic remains.

(5) The *Protatrypa lepidota* oryctocoenosis (Bed 66/25) is enclosed into foliated limestones. The structure is homogeneous. Subautochthonous organic remains are small.

(6) The *Beyrichia kureikiana* oryctocoenosis (beds 66/26–66/30) is enclosed into bedded and cloddish limestones as medium-sized accumulations. In the main oryctoparcella (platy limestones), the density of subautochthonous burials of medium-sized fossils is low. In subordinate oryctoparcellas (cloddish limestones), concentration of small autochthonous organic remains is also low. Gastropods are predominant in the first case, and ostracodes in the second.

(7) The *Stelodictyon moierense* oryctocoenosis (Bed 66/31) from biohermal limestones is homogeneous in structure. The density of burials is low. Autochthonous fossils are large, and stromatolites are abundant among them.

The lower Tukal Subhorizon is rich in stromatolites, ostracodes, brachiopods and crinoids. Tabulates, stromatoporoids, and rugose corals are less frequent, and other fossils occur here as single specimens.

(1) The *Stromatolithi* oryctocoenosis (beds 11/3–11/10) is enclosed into limestone beds and lenses. The structure is banded or of the small-accumulation type. The moderate density of burials is characteristic of certain levels of the main oryctoparcella (limestone beds), where dominant fossils are ostracodes or stromatolites. Subautochthonous organic remains are small. Large stromatolite calyptras often form biostromes. In subordinate oryctoparcellas, fossils are occasional and small.

(2) The *Eukloedenella kureikensis* oryctocoenosis (beds 11/11–11/13) from marl beds and limestone lentils is of the small-cluster type. Marl beds yield rare and small organic remains. Subordinate oryctoparcellas enclose either abundant subautochthonous shells of ostracodes and gastropods (lentils of brecciated limestone), or dispersed allochthonous shells of the same taxonomic groups.

(3) The *Stromatolithi* oryctocoenosis (beds 11/14–11/17) from platy limestones and stromatolitic biostromes is of the large-cluster type. In platy limstones with low density of burials, organic remains are small. Stromatolitic biostromes are extended, although the solitary calyptras of medium size are also frequent. Around the biostromes and calyptras, shells of *Hyattidina acutisummitatus* are often buried in the intravital state.

(4) The *Hyattidina acutisummitatus* oryctocoenosis (beds 3/2–3/12) is enclosed into marl beds with limestone clods and in lentils of platy and cloddish limestones. The main oryctoparcella (marl beds) is characterized by low density of autochthonous burials of small fossils. Subautochthonous or allochthonous burials of the same organic remains are concentrated in subordinate oryctoparcellas (lentils of platy limestone). Abundance of fossils in autochthonous burials (lentils of cloddish limestone) is low.

(5) The *Subalveolites subulosus* oryctocoenosis (Bed 3/13) from massive limestones is homogeneous in structure. The density of burials is high. Autochthonous organic remains are medium in size.

(6) The *Bystrowicrinus quinquelobatus* oryctocoenosis (beds 3/14–3/28) is enclosed into platy limestones, biostromes, and marl lentils. Its structure is of the small-accumulation type. The high density of autochthonous and subautochthonous burials of small organic remains is typical of the main oryctoparcella (limestones). The medium-sized biostromes are composed of stromatolites. Subordinate oryctoparcellas (marl lentils) display the low concentration of small subautochthonous fossils.

Widespread in the upper Tukal Subhorizon are ostracodes, brachiopods, crinoids, and gastropods; less abundant are stromatoporoids, tabulates, and bryozoans. Other fossils occur sporadically.

(1) The *Hyattidina acutisummitatus* oryctocoenosis (beds 3/29-3, 2/4–2/15) from marls enclosing the limestone clods and lentils is of the small-cluster structure. In the main oryctoparcella (marls), rare subautochthonous burials consist of small organic remains. Subordinate oryctoparcellas (limestone lentils) contain autochthonous and allochthonous burials of highly concentrated small organic remains.

(2) The *Bazaricrinus parvulus* oryctocoenosis (beds 2/16–2/18, 1/1–1/9) is confined to platy limestones with marl lentils. Its structure is composed by small and medium-sized fossil accumulations. The moderate density of subautochthonous burials of small organic remains is characteristic of the main oryctoparcella (platy limestones). Subordinate marl lentils enclose autochthonous to subautochthonous burials, where concentration of the same fossils is low.

(3) The *Sibiricrinus helenae* oryctocoenosis (beds 1/10–1/14) of the small-cluster structure characterizes marl beds enclosing limestone clods and lentils. The principal oryctoparcella (marls) shows the moderate density of subautochthonous burials of small fossils. In autochthonous and allochthonous burials of subordinate oryctoparcellas (limestone lentils), concentration of the same organic remains is either high or low.

(4) The *Prosolarium cirrhosa* oryctocoenosis (beds 1/15, 1/16, 4/2, 4/3) from bedded limestones with marl intercalations is of the homogeneous structure in general. The density of burials is moderate to high in limestones and low in marls. Autochthonous to subautochthonous organic remains are small.

(5) The *Eukloedenella kureikensis* oryctocoenosis (beds 4/4–4/6) of the small-cluster structure is from marls enclosing clods and small lenses of limestone. In the main oryctoparcella (marls), the density of autochthonous burials of small organic remains is low. The same fossils are more concentrated in autochthonous and allochthonous burials of subordinate oryctoparcellas (limestone lenses).

(6) The *Leperditia lumaea* oryctocoenosis (beds 4/7–4/12) is enclosed into beds of limestone, dolomitic marl, and foliated dolomite. Clusters of fossils vary in size from small to large. Marls and some lentils of dolomite and dolomitic limestone are nearly barren of fossils. Remains of *Densastroma astroites* are concentrated in limestone lentils of Bed 12. Other small organic remains (frequent bryozoans in association with rare ostracodes and crinoids) occur in subautochthonous burials within limestone lentils.

(7) The *Murchsonia cingulata* oryctocoenosis (Bed 4/13) from black massive limestones is homogeneous in structure. The density of burials is very high. Autochthonous to subautochthonous organic remains are medium-sized. Shells of diverse gastropods and cephalopods are enveloped into phytogenic films. Concentration of meso- and microphytoliths is high. The small-shelled fauna is confined to subautochthonous burials.

(8) The *Eohowellella minima* oryctocoenosis (Bed 4/14) is from marl beds enclosing clods and rare lentils of limestones. It is of the small-cluster structure. Marls are almost barren of fossils. In subordinate oryctoparcellas (limestone lentils), density of subautochthonous burials of small shells is low.

(9) The *Murchsonia cingulata* oryctocoenosis (beds 10/28–10/31) from black platy limestones and biostromes is of the medium-cluster structure. In the principal oryctoparcella (limestone beds), density of autochthonous burials of small and medium-sized fossils is moderate. Stromatolitic biostromes represent subordinate oryctoparcellas.

stones) generally bear the medium-sized stromatolites and stromatoporoids in frequent autochthonous burials.

(5) The *Favososphaeridium polybrochatum* oryctocoenosis (beds 58A/5–58A/6) is enclosed into dolomitic marls. The rocks bear only rare acritarchs.

(6) The *Beyrichia mirabilis* oryctocoenosis (beds 58A/7–58A/8, 58/2–58/14) is enclosed into marls intercalated with limestone lentils. The structure is banded. The low density of burials of small autochthonous fossils is characteristic of the main oryctoparcella (marls). Subordinate oryctoparcellas (limestone lentils) enclose rather frequent subautochthonous organic remains medium in size.

(7) The *Straparollus alacer* oryctocoenosis (Bed 58/15) from black and gray limestones is of the small-cluster structure. The density of burials of medium-sized autochthonous fossils is very high in the main oryctoparcella (black limestones). Subordinate oryctoparcella (gray limestones) is of a high burial density and consists of stromatolites.

(8) The *Beyrichia mirabilis* oryctocoenosis (beds 58/16–58/19) is enclosed into limestone and marls displaying wavy lamination. The structure is of the local small-accumulation type. The main oryctoparcella (limestones) shows the low density of subautochthonous burials of small fossils. Fossils of subordinate oryctoparcellas (marl interbeds) are represented by acritarchs only.

(9) The *Dalejina ribnayaensis* oryctocoenosis (Bed 58/20) is enclosed into marls with limestone lentils. The structure is of the small-accumulation type. The main oryctoparcella (marls) shows the low density of subautochthonous burials of small fossils. The same fossils occur in allochthonous position in subordinate oryctoparcellas (limestones).

(10) The *Dalejina ribnayaensis* oryctocoenosis from foliated limestones (beds 58/21–58/22) is homogeneous in structure. It consists of small brachiopod shells only.

(11) The *Labechia condensa* oryctocoenosis (beds 64/12–64/18) from platy limestones and biostromes is of the medium-cluster type. The main oryctoparcella (bedded limestones) shows the moderate density of autochthonous burials of small organic remains. Subordinate oryctoparcellas (biostromes) are rich in autochthonous stromatoporoids, stromatolites, and tabulates of medium size.

The upper Khakoma Subhorizon is found to be rich in stromatoporoids, tabulates, crinoids, and gastropods. These fossils coexist with less abundant cephalopods, bryozoans, brachiopods, and sporadically with singular specimens of other taxa.

(1) The *Yavorskiina membrosa* oryctocoenosis (beds 66/1b–66/8) is enclosed into platy limestones and bioherms. Its structure is of the large-cluster type. The main oryctoparcella (platy limestones) bears rare autochthonous organic remains of small and medium size. Subordinate oryctoparcellas (bioherms) enclose more frequent allochthonous and autochthonous burials of large stromatoporoids, tabulates, rugose corals, and crinoids.

(2) The *Bystrowicrinus bilobatus* oryctocoenosis (beds 66/9–66/12) is enclosed into massive and platy limestones. It is of the small-accumulation structure. The main oryctoparcella (massive limestones) shows the low density of allochthonous burials of small organic remains and abundant calyptras of stromatolites. Rare small-shelled fauna is concentrated in subordinate oryctoparcellas (platy limestones).

(3) The *Mudeaters* oryctocoenosis (beds 66/13–66/15) is enclosed into cloddish, platy, and biostromal limestones. Its structure is banded. Cloddish limestones yield rare and small brachiopod shells only. *Clavidictyon cylindricum* is dominant species in biostromes, where the burial density of medium-sized organic remains is high.

(4) The *Clavidictyon cylindricum* oryctocoenosis (beds 66/19–66/24) is enclosed into foliated and platy limestones and marls in a form of small accumulations and small clusters of fossils. The foliated limestones are almost barren of fossils and bear only rare crinoid columnals. Abundant autochthonous burials in marls and biostromes consist of medium-sized organic remains.

(5) The *Protatrypa lepidota* oryctocoenosis (Bed 66/25) is enclosed into foliated limestones. The structure is homogeneous. Subautochthonous organic remains are small.

(6) The *Beyrichia kureikiana* oryctocoenosis (beds 66/26–66/30) is enclosed into bedded and cloddish limestones as medium-sized accumulations. In the main oryctoparcella (platy limestones), the density of subautochthonous burials of medium-sized fossils is low. In subordinate oryctoparcellas (cloddish limestones), concentration of small autochthonous organic remains is also low. Gastropods are predominant in the first case, and ostracodes in the second.

(7) The *Stelodictyon moierense* oryctocoenosis (Bed 66/31) from biohermal limestones is homogeneous in structure. The density of burials is low. Autochthonous fossils are large, and stromatolites are abundant among them.

The lower Tukal Subhorizon is rich in stromatolites, ostracodes, brachiopods and crinoids. Tabulates, stromatoporoids, and rugose corals are less frequent, and other fossils occur here as single specimens.

(1) The *Stromatolithi* oryctocoenosis (beds 11/3–11/10) is enclosed into limestone beds and lenses. The structure is banded or of the small-accumulation type. The moderate density of burials is characteristic of certain levels of the main oryctoparcella (limestone beds), where dominant fossils are ostracodes or stromatolites. Subautochthonous organic remains are small. Large stromatolite calyptras often form biostromes. In subordinate oryctoparcellas, fossils are occasional and small.

(2) The *Eukloedenella kureikensis* oryctocoenosis (beds 11/11–11/13) from marl beds and limestone lentils is of the small-cluster type. Marl beds yield rare and small organic remains. Subordinate oryctoparcellas enclose either abundant subautochthonous shells of ostracodes and gastropods (lentils of brecciated limestone), or dispersed allochthonous shells of the same taxonomic groups.

(3) The *Stromatolithi* oryctocoenosis (beds 11/14–11/17) from platy limestones and stromatolitic biostromes is of the large-cluster type. In platy limstones with low density of burials, organic remains are small. Stromatolitic biostromes are extended, although the solitary calyptras of medium size are also frequent. Around the biostromes and calyptras, shells of *Hyattidina acutisummitatus* are often buried in the intravital state.

(4) The *Hyattidina acutisummitatus* oryctocoenosis (beds 3/2–3/12) is enclosed into marl beds with limestone clods and in lentils of platy and cloddish limestones. The main oryctoparcella (marl beds) is characterized by low density of autochthonous burials of small fossils. Subautochthonous or allochthonous burials of the same organic remains are concentrated in subordinate oryctoparcellas (lentils of platy limestone). Abundance of fossils in autochthonous burials (lentils of cloddish limestone) is low.

(5) The *Subalveolites subulosus* oryctocoenosis (Bed 3/13) from massive limestones is homogeneous in structure. The density of burials is high. Autochthonous organic remains are medium in size.

(6) The *Bystrowicrinus quinquelobatus* oryctocoenosis (beds 3/14–3/28) is enclosed into platy limestones, biostromes, and marl lentils. Its structure is of the small-accumulation type. The high density of autochthonous and subautochthonous burials of small organic remains is typical of the main oryctoparcella (limestones). The medium-sized biostromes are composed of stromatolites. Subordinate oryctoparcellas (marl lentils) display the low concentration of small subautochthonous fossils.

Widespread in the upper Tukal Subhorizon are ostracodes, brachiopods, crinoids, and gastropods; less abundant are stromatoporoids, tabulates, and bryozoans. Other fossils occur sporadically.

(1) The *Hyattidina acutisummitatus* oryctocoenosis (beds 3/29-3, 2/4–2/15) from marls enclosing the limestone clods and lentils is of the small-cluster structure. In the main oryctoparcella (marls), rare subautochthonous burials consist of small organic remains. Subordinate oryctoparcellas (limestone lentils) contain autochthonous and allochthonous burials of highly concentrated small organic remains.

(2) The *Bazaricrinus parvulus* oryctocoenosis (beds 2/16–2/18, 1/1–1/9) is confined to platy limestones with marl lentils. Its structure is composed by small and medium-sized fossil accumulations. The moderate density of subautochthonous burials of small organic remains is characteristic of the main oryctoparcella (platy limestones). Subordinate marl lentils enclose autochthonous to subautochthonous burials, where concentration of the same fossils is low.

(3) The *Sibiricrinus helenae* oryctocoenosis (beds 1/10–1/14) of the small-cluster structure characterizes marl beds enclosing limestone clods and lentils. The principal oryctoparcella (marls) shows the moderate density of subautochthonous burials of small fossils. In autochthonous and allochthonous burials of subordinate oryctoparcellas (limestone lentils), concentration of the same organic remains is either high or low.

(4) The *Prosolarium cirrhosa* oryctocoenosis (beds 1/15, 1/16, 4/2, 4/3) from bedded limestones with marl intercalations is of the homogeneous structure in general. The density of burials is moderate to high in limestones and low in marls. Autochthonous to subautochthonous organic remains are small.

(5) The *Eukloedenella kureikensis* oryctocoenosis (beds 4/4–4/6) of the small-cluster structure is from marls enclosing clods and small lenses of limestone. In the main oryctoparcella (marls), the density of autochthonous burials of small organic remains is low. The same fossils are more concentrated in autochthonous and allochthonous burials of subordinate oryctoparcellas (limestone lenses).

(6) The *Leperditia lumaea* oryctocoenosis (beds 4/7–4/12) is enclosed into beds of limestone, dolomitic marl, and foliated dolomite. Clusters of fossils vary in size from small to large. Marls and some lentils of dolomite and dolomitic limestone are nearly barren of fossils. Remains of *Densastroma astroites* are concentrated in limestone lentils of Bed 12. Other small organic remains (frequent bryozoans in association with rare ostracodes and crinoids) occur in subautochthonous burials within limestone lentils.

(7) The *Murchsonia cingulata* oryctocoenosis (Bed 4/13) from black massive limestones is homogeneous in structure. The density of burials is very high. Autochthonous to subautochthonous organic remains are medium-sized. Shells of diverse gastropods and cephalopods are enveloped into phytogenic films. Concentration of meso- and microphytoliths is high. The small-shelled fauna is confined to subautochthonous burials.

(8) The *Eohowellella minima* oryctocoenosis (Bed 4/14) is from marl beds enclosing clods and rare lentils of limestones. It is of the small-cluster structure. Marls are almost barren of fossils. In subordinate oryctoparcellas (limestone lentils), density of subautochthonous burials of small shells is low.

(9) The *Murchsonia cingulata* oryctocoenosis (beds 10/28–10/31) from black platy limestones and biostromes is of the medium-cluster structure. In the principal oryctoparcella (limestone beds), density of autochthonous burials of small and medium-sized fossils is moderate. Stromatolitic biostromes represent subordinate oryctoparcellas.

(10) The *Parastriatopora kureikiana* oryctocoenosis (beds 10/32–10/36) characterizes platy dolomites, marls, limestone lentils, and biostromes. Its mosaic structure is of the small-cluster type. Most typical amid variable oryctoparcellas are those enclosing the subautochthonous mass accumulations of small ramose tabulates. Beds of foliated dolomite are impoverished in organic remains. Mass accumulations of bryozoans are confined to brecciated limestone lentils. Marls enclosing limestone lentils are almost barren of fossils. Rare stromatolitic biostromes are locally rich in tabulates and stromatoporoids.

(11) The *Leperditia lumaea* oryctocoenosis (beds 10/37–10/40) from marl, platy limestone, and dolomite sequence is of the banded structure. Despite the difference in lithology of beds, the oryctocoenosis is rather homogeneous. The density of burials is low. The burials are mainly of the subautochthonous type and consist of small organic remains.

The stratotype of the Postnichnyi Horizon is barren of fossils. The section of the Middendorf Cave locality at the Nizhnyaya Taimyra River (Exposure 218) is chosen as the ecological and paleontological standard for the Ludfordian and Pridoli of East Siberia. This section has been characterized in detail earlier (Tesakov *et al.*, 1995).

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Abstract—The structure of the Jurassic deposits of the Episkopi Formation from southwestern Cyprus is considered. The section is composed (from the base upward) of (1) the Callovian–Oxfordian siliceous rocks and turbidites (30 m) with rich radiolarian assemblages, (2) the Oxfordian mudstone (25 m) with poor radiolarian assemblage, and (3) the Kimmeridgian–Tithonian mudstone and turbidite (10 m) with rich radiolarian assemblages. The radiolarian assemblages are correlated with synchronous faunas from the West Mediterranean region and eastern Russia. The character of radiolarian assemblages from the Episkopi Formation depends on types of sediments. The registered stratigraphic interval (Oxfordian), during which the assemblage was impoverished, coincides with the time of intensified terrigenous influx that was accompanied by disappearance of coarse-grained turbidites. Thus, during the Jurassic sedimentogenesis in Cyprus, the gradual intensification of the terrigenous non-turbidite sedimentation was combined with periods of growing silica accumulation and formation of turbidites.

Key words: *Cyprus, stratigraphy, Jurassic, radiolarians, lithology, turbidites, silica accumulation.*

The complicated Mamonia allochthon of southwestern Cyprus consists of a series of tectonic slices, the upper of which is composed of rocks of the Ayios Photios Group (Krylov *et al.*, 1993). The group comprises the Episkopi Formation spanning the interval from the Callovian to Cenomanian and representing the most complete section of the Upper Jurassic–Lower Cretaceous depposits in the region. The study of sedimentation conditions and stratigraphy of the formation is of basic importance for deciphering and understanding the Cenozoic history of southern Mesotethys. First, the continuous succession of radiolarian assemblages from the Callovian to Neocomian is distinguished here (Bragin and Krylov, 1991) as comparable with biostratigraphic units known in various regions of the world, including eastern Russia (Bragin, 1991, 1993; Vishnevskaya, 1991). Second, the study of this material provides the possibility to correlate the evolution of the East Mediterranean Mesozoic basin with other paleooceans, for instance, with the northwestern Pacific, using data on radiolarian assemblages and lithology of sediments. The understanding of stratigraphy in detail is important for more substantiated correlation of geological events in the Mesozoic history, to which an active interest has already been shown (Knipper and Sharaskin, 1995; Zonenshain, 1987; Kaz'min *et al.*, 1987; Robertson, 1990; Robertson and Woodcock, 1979). It should be noted that tectonic reconstructions are usually based on weak stratigraphic evidence acquired from the Dhiarozos Series intensely dislocated and yet imperfectly understood (Sharas'kin *et al.*, 1995).

Our study is of a special significance after revealing a remarkable hiatus between the Triassic Vlambouros and Jurassic Episkopi formations (Bragin and Krylov, 1991; Bragyn and Krylov, 1996). At the base of the Episkopi Formation above the Upper Triassic rocks of the Vlambouros Formation, there is a stratigraphic hiatus that is not accompanied by an angular unconformity, being manifested as the absence of the Lower and most of the Middle Jurassic deposits in the section. Basal beds of the formation are most likely of the Callovian–Oxfordian age. The study of deposits from the Episkopi Formation is of prime interest for solving various regional problems as well as for some methodical issues.

We studied the Episkopi Formation in a variety of sections, none of which spans the whole stratigraphic interval of the discussed unit (Figs. 1, 2). To understand the structure of the formation, all particular sections have to be considered. The description of three key sections is given below.

In the extreme west, within the Lara Bay (Akamas Peninsula), the Upper Triassic sandstone of the Vlambouros Formation is overlapped by a sequence of the following units (Figs. 1, 2).

(1) Red, cherry-red, rarely light green, platy cherts; the rocks bear fragments of radiolarians *Acanthocircus suboblongus* (Yao), *A.* sp. cf. *A. trizonalis* (Rust), *Amphipyndax* (?) sp. aff. *A. tsunoensis* (Aita), *Angulobracchia purisimaensis* (Pessagno), *Andromeda* sp. cf. *A. podbielensis* (Ozvoldova), *Archaeodictyomitra* sp. cf. *A. rigida* Pessagno, *Guexella nudata* (Kocher),

(10) The *Parastriatopora kureikiana* oryctocoenosis (beds 10/32–10/36) characterizes platy dolomites, marls, limestone lentils, and biostromes. Its mosaic structure is of the small-cluster type. Most typical amid variable oryctoparcellas are those enclosing the subautochthonous mass accumulations of small ramose tabulates. Beds of foliated dolomite are impoverished in organic remains. Mass accumulations of bryozoans are confined to brecciated limestone lentils. Marls enclosing limestone lentils are almost barren of fossils. Rare stromatolitic biostromes are locally rich in tabulates and stromatoporoids.

(11) The *Leperditia lumaea* oryctocoenosis (beds 10/37–10/40) from marl, platy limestone, and dolomite sequence is of the banded structure. Despite the difference in lithology of beds, the oryctocoenosis is rather homogeneous. The density of burials is low. The burials are mainly of the subautochthonous type and consist of small organic remains.

The stratotype of the Postnichnyi Horizon is barren of fossils. The section of the Middendorf Cave locality at the Nizhnyaya Taimyra River (Exposure 218) is chosen as the ecological and paleontological standard for the Ludfordian and Přidoli of East Siberia. This section has been characterized in detail earlier (Tesakov *et al.*, 1995).

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Reviewer B.S. Sokolov

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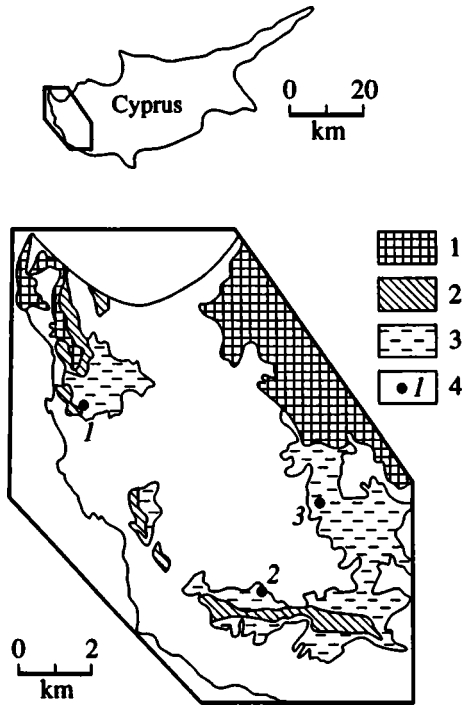

Fig. 1. Location of key sections of the Jurassic deposits of the Episkopi Formation: (1) ophiolites of Troodos massif; (2) terrigenous and ophiolitic melanges; (3) the Ayios Photios Group; (4) section localities (*1*, Lara Bay; *2*, Episkopi Village; *3*, Ayios Photios Village).

Higumastra imbricata (Ozvoldova), *H.* sp. cf. *H. inflata* Baumgartner, *Hsuum brevicostatum* Ozvoldova, *H. maxwelli* Pessagno, *H.* sp., *Mirifusus* sp. cf. *M. guadalupensis* Pessagno, *M. dianae* (Karrer), *Paronaella mulleri* Pessagno, *Parvicingula* sp. cf. *P. dhimenaensis* Baumgartner, *Protunuma ochiensis* Matsuoka, *P. turbo* Matsuoka, *Ristola altissima* (Rust), *Semihsuum sourdoughense* Pessagno, Blome et Hull, *Stichocapsa robusta* Matsuoka, *S. convexa* Yao, *S.* sp. aff. *S. himedaruma* Aita, *Theocapsa* sp. cf. *T. pentagona* Aita, *Tricolocapsa* sp. cf. *T. parvipora* Tan, *Tritrabs ewingi* (Pessagno), presumably of the Callovian–early Oxfordian age. The thickness is 12 m.

(2) Red-brown, coarse-pebbly conglomerates with calcareous cement and pebbles of radiolarian chert, siliceous mudstone and micrite; the rocks enclose interlayers of cross-laminated calcarenite. Pebbles are rounded to a different extent. Moreover, the unit comprises fragments of radiolarite beds. Radiolarites contain remains of Jurassic radiolarians similar to those from Bed 1. The thickness is 4 m.

(3) Light gray and greenish cherts with interlayers of red cherts and siliceous green mudstone. Cherts yielded radiolarians *Acanthocircus suboblongus* Yao, *Archaeodictyomitra rigida* Pessagno, *Hsuum brevicostatum* Ozvoldova, *Hsuum* sp., *Stichocapsa robusta* Matsuoka, *S. himedaruma* Aita, *Stichomitra tairai* Aita, *Stylocapsa oblongula* Kocher, and *Theocapsomma cordis* Kocher, probably of the Oxfordian age. The thickness is 10 m.

(4) Red–brown conglomerates with calcareous cement, grading upward into cross-laminated calcarenites and banded green cherts enclosing the radiolarian assemblage similar to that from Bed 3. The thickness is 3.5 m.

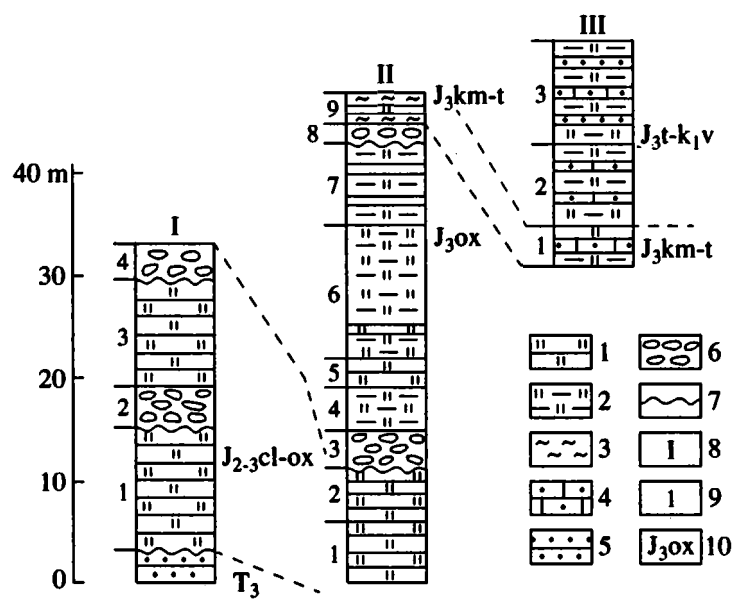

Fig. 2. Structure and correlation of sections of the Jurassic Episkopi Formation: (1) cherts; (2) siliceous mudstone; (3) clay; (4) calcarenite; (5) sandstone; (6) conglomerate and gravelstone; (7) unconformity; (8) section number (I, Lara; II, Episkopi; III, Ayios Photios); (9) bed number; (10) age.

Similar deposits (only without coarse conglomerates) are exposed in lower reaches of the Mavrokolymbos River, where calcarenite interlayers contain benthic foraminifers *Ammobaculites* sp. cf. *A. braunsteini* Cushm. et Applin, *Haplophragmium* sp., *Haurania* sp. cf. *H. desepta* Hens., *Kurnubia* sp., *Mayncina termieri* Hottin., *Orbitopsella* sp., *Pfenderina* sp., *Protopeneroplis striata* Weyn., and *Valvulina lugeoni* Sept. of the Middle Jurassic age (identifications by K.I. Kuznetsova). Radiolarians presumably of the Callovian–early Oxfordian age similar to those from the section near Lara Bay are also found in this locality. Foraminifers are certainly redeposited. This assumption can be supported by the following arguments. First, foraminifers are confined to calcarenite interlayers with abundant oolites, pellets, shell detritus, i.e., with carbonate material of the shallow-water origin. Second, foraminifers are represented by benthic species of a relatively shallow-water origin.

The lower part of the Episkopi Formation is also exposed in the stratotype section at the left slope of the Ezousa River valley opposite to the northern margin of Episkopi Village (Figs. 1, 2). The contact with the Vlambouros Formation is not exposed here. The succession of units from the base upward is as follows.

(1) Red, pink, more rarely greenish and light gray, platy, in places clayey cherts with interlayers of pink and green, silicified calcarenites include the cherty conglomerate horizon (0.5 m) in the upper part of the bed. The visible thickness is 6 m.

(2) Alternation of radiolarian cherts, similar to those from the previous bed, with red and crimson-colored, more rarely greenish gray pelitic cherts; the unit encloses two beds (up to 0.5 m) of cherty conglomerates at the base and bears, as the previous one, radiolarians presumably of the Callovian–early Oxfordian age. Radiolarians are similar to those found in the Lara Bay section (Bed 1). The thickness is 5 m.

(3) Coarse conglomerates and brown–gray to brown massive gravelstones with abundant detritus and fragments of redeposited chert and calcarenite beds. The lower contact of the unit is abrupt and uneven. The calcarenite–calcirudite horizon (1 m) is located in the middle part. The thickness is 4 m.

(4) Mudstone, siliceous, brick-red and reddish–brown, loose, with red clay interlayers; the thickness is 4 m.

(5) Silicified, greenish gray cherts and calcarenites with interlayers of siliceous, greenish gray mudstone; the thickness is 3 m.

(6) Siliceous, reddish-brown and greenish-gray mudstone with frequent interlayers of pelitic, reddish–brown cherts in the lower part and with red clay interlayers in the upper part; the thickness is 13 m.

(7) Alternation of greenish gray siliceous pelites, silicified greenish gray and reddish calcarenites, greenish gray radiolarites and siliceous, and greenish gray to reddish brown mudstones; units 4–7 are characterized by the radiolarian assemblage of *Acanthocircus suboblongus* Yao, *Hsuum brevicostatum* Ozvoldova, *Protunuma ochiensis* Matsuoka, *Stylocapsa oblongula* (Kocher), *Stichocapsa himedaruma* Aita, and *Theocapsomma cordis* Kocher, presumably of the Oxfordian age. The thickness is 8 m.

(8) Grayish brown, cherty and carbonate-detrital lumpy gravelstones with fragments of redeposited chert and calcarenite beds. The lower contact is very uneven, truncating underlying beds and filled pockets. The thickness is 2 m.

(9) Reddish brown clay with interlayers of greenish gray and white cherts bearing radiolarians *Podocapsa amphitreptera* Foreman and *Spongocapsula perampla* (Rust) presumably of the Kimmeridgian–Tithonian age. The visible thickness is 3 m.

Units 1–3 of the section correspond to the section interval exposed near the Lara Bay. These beds represent the lower, mainly siliceous member of the Episkopi Formation. They enclose horizons of intraformational conglomerates found in other areas as well (except for the Hapotami River basin). Units 4–8 of the stratotype section belong to the middle, predominantly clayey member of the formation. Unit 9 of the stratotype section is attributed to the upper part of the Episkopi Formation studied in the section near Ayios Photios Village (Figs. 1, 2).

Here, along the road from Ayios Photios to Phalia Village, the section exposed is represented (from the base upward) by three units.

(1) Alternation of light green–gray and pinkish green calcarenites, in places banded and often silicified, with greenish gray and red radiolarian cherts and brown–crimson siliceous mudstone. The unit yields the Kimmeridgian–Tithonian radiolarians *Acaeniotyle diaphorogona* Foreman, *Alievium* sp., *Archaeospongoprunum* sp. cf. *A. patricki* Jud, *Emiluvia orea* Baumgartner, *E. pessagnoi* Foreman, *E. premyogii* Baumgartner, *Lychnocanomma* sp. cf. *L. xiphophora* (Rust), *Mirifusus dianae* (Karrer), *Paronaella* sp. aff. *P. kotura* Baumgartner, *P.* sp. ex gr. *P. pristidentata* Baumgartner, *Podobursa tricantha* (Fischli), *Podocapsa amphitreptera* Foreman, *Pseudocrucella adriani* Baumgartner, *Pseudoeucyrtis* sp., *Ristola altissima* (Rust), *Spongocapsula perampla* (Rust), *S.* sp. Radiolarians *Obesacapsula verbana*, (Parona) and *Sethocapsa cetia* Foreman appear in the upper part of the bed. The visible thickness is 4 m.

(2) Siliceous, brown–crimson thin-platy mudstone and siltstone with horizons (less than 0.05 m) of light greenish gray banded calcarenites; the rocks bear radiolarians *Archaeodictyomitra apiara* (Rust), *Alievium helenae* Schaaf, *Ditrabs sansalvadorensis* (Pessagno), *Emiluvia pessagnoi* Foreman, *Mirifusus mediodilatatus* (Rust), *Parvicingula boesii* (Parona), *Podobursa polylophia* Foreman, *Podocapsa amphitreptera* Foreman, *Pseudodictyomitra carpatica* (Lozyniak), *Xitus*

sp. aff. *X. alievi* (Foreman), most likely of the Tithonian–Valanginian age. The thickness is 8 m.

(3) Alternation of siliceous brown–crimson thin-platy conchoidal mudstone and siltstone with gray and greenish gray medium-grained massive quartz–feldspathic sandstone displaying graded bedding (the thickness of interlayers is up to 0.2 m), and with light greenish gray banded calcarenites and pink–green compact conchoidal marls. Quartz–feldspathic sandstone differs in composition from that of the Triassic Vlambouros Formation and is similar to massive sandstone found in tectonic blocks in the Akamas Peninsula. The radiolarian assemblage is similar to that from the previous unit. The thickness is 10 m.

The last unit is overlain by mudstone with calcarenite interlayers containing the Valanginian–Hauterivian radiolarians. This allows us to date the Akamas sandstone and to regard them as an element of the Episkopi Formation.

Thus, the Jurassic deposits of the Episkopi Formation represent a thick (up to 80–90 m, considering all the sections presented in the paper, excluding the overlap) complex of different facies, which is of a complicated structure (Fig. 2). The section is subdivided into members as follows.

Member 1: radiolarian cherts presumably of the Callovian–Oxfordian age (up to 30 m).

Member 2: Siliceous mudstones presumably of the Oxfordian age (about 25 m).

Member 3: Alternating siliceous mudstone, sandstone, calcarenite and radiolarite, most likely of the Kimmeridgian–Valanginian age (up to 25 m).

Member 1 shows the fine rhythmical alternation of extended radiolarite beds up to 5–7 m thick with siliceous mudstones (discontinuous beds up to 2–3 m thick). Calcilutite interlayers are few and far between. Horizons of coarse, calcareous–siliceous conglomerates are also found. They are formed due to destruction and washout of accumulating sediments. Rich radiolarian assemblages occur everywhere. The sediments appear to be accumulated in a deep, below the carbonate compensation depth (CCD), basin with calm sedimentation conditions violated only by rare interference of distal calciturbidites and sometimes interrupted by redeposition with formation of intraformational conglomerates. During formation of the conglomerates, radiolarites were slightly lithified that resulted in the appearance of deformed fragments of radiolarite beds among pebbles of siliceous rocks. The carbonate matrix did not seem to be lithified at all. These structural peculiarities could reflect the sequences formation on a slope and the respective detachment, sliding and the redeposition of the member fragments. The deposits are similar in the type of microfauna with synchronous sequences of the West Mediterranean region (Baumgartner, 1984; Aita, 1987; Baumgartner *et al.*, 1995). Two radiolarian assemblages are distinguished here (the table; Plates I, II): one of presumably Callovian–Oxfordian age with *Gueuxella nudata, Protunuma turbo*, and *Theocapsomma* sp. cf. *T. pentagona* and another of presumably Oxfordian age with *Stylocapsa oblongula, Stichomitra tairai, Theocapsomma* sp. cf. *T. constricta* and other forms. Both assemblages include many transitional species. The lower assemblage correlates with radiolarians of the Upper Callovian–Lower Oxfordian *Gueuxella nudata–Archaeodictyomitra (?) mirabilis* and *Amphipyndax tsunoensis* zones of Japan and Italy (Aita, 1987), as well as with the Callovian–Oxfordian assemblages of tropical areas of the Atlantic and West Mediterranean regions, and with the Callovian–lower Oxfordian Zone A-2 (Baumgartner, 1984). It should be pointed out that the most recent data on the Mediterranean region (Baumgartner *et al.*, 1995) suggest the more extended vertical distribution (beginning with the Bajocian) of many species of the assemblage (*Gueuxella nudata, Protunuma ochiensis, Mirifusus guadalupensis, Higumastra imbricata*). However, forms such as *Unuma echinatus, U. typicus* and others typical of the Bajocian–Bathonian, have not been found in the assemblage studied. At the same time, the *Semihsuum sourdoughense* and *Hsuum maxwelli* typical of the Oxfordian deposits in North America (Pessagno *et al.*, 1993) are present. It should be emphasized that the pictured *Hsuum maxwelli* from the Mediterranean region (Baumgartner, 1984; Baumgartner *et al.*, 1995) does not correspond to the species diagnosis (Pessagno *et al.*, 1993). Just for these reasons, we attribute our assemblage to the Callovian–Oxfordian.

The upper assemblage correlates with those from the Oxfordian (probably middle) *Amphipyndax tsunoensis–Stylocapsa (?) spiralis* and *Gongylothorax sakawaensis* zones of Italy and Japan (Aita, 1987), as well as with assemblages of the Callovian–Oxfordian zones (Baumgartner, 1984; Baumgartner *et al.*, 1995). Species of the wide stratigraphic range are also present in this interval (Baumgartner *et al.*, 1995), however, by the position in the section it is assumed to be of the Oxfordian age. The assemblage composition indicates conditions of an open marine basin where the sequence formed. However, shallow-water areas of the carbonate shelf were located close to the region of siliceous mud accumulation. The areas served as a source of carbonate material of calciturbidites and conglomerates, and redeposition of benthic foraminifers therewith took place as well. All this indicates the washout of shelf deposits. Calciturbidites also include terrigenous clastics, though in a lesser amount. This material is usually well-rounded and suggests that it was a supracrustal rock complex subjected to erosion. The presence of horizons with signs of sliding of unconsolidated sediments in some sections shows that sediments accumulated on the bottom slightly sloping at an angle of less than 1°. The sedimentation rate for siliceous sediments was higher than that for the Triassic siliceous sequences in northeastern Asia, which are assumed to be of the oceanic origin (Bragin, 1991).

Distribution of radiolarian species in the section of the Jurassic Episkopi Formation

Radiolarian species	Member 1 (Callovian–Oxfordian)	Member 2 (Oxfordian)	Member 3 (Kimmeridgian–Valanginian)
A. suboblongus	++++++++++	++++++++++	
A. purisimaensis	+++++		
G. nudata	+++++		
H. imbricata	+++++		
H. brevicostatum	++++++++++	++++++++++	
H. maxwelli	+++++		
M. dianae	++++++++++	++++++++++	++++++++++++++
P. ochiensis	++++++++++	++++++++++	
R. altissima	++++++++++	++++++++++	+++++++++
S. sourdoughense	+++++		
S. robusta	++++++++++		
S. convexa	+++++		
T. ewingi	+++++		
S. himedaruma	+++++	++++++++++	
S. tairai	+++++		
S. oblongula	+++++	++++++++++	
T. cordis	+++++	++++++++++	
A. diaphorogona			++++++++++++++
E. orea			+++++++
E. pessagnoi			++++++++++++++
E. premyogii			+++++++
P. triacantha			++++++++++++++
P. amphitreptera			++++++++++++++
P. adriani			+++++++
S. perampla			+++++++
O. verbana			++
S. cetia			++++++++
A. apiara			++++++
A. helenae			++++++
D. sansalvadorensis			++++++
P. boesii			++++++
P. polylophia			++++++
P. carpatica			++++++

Dominant in Member 2 are siliceous mudstones usually red in color and representing pelitic background sediments bearing rare and isolated radiolarian remains and lacking carbonate or terrigenous sandy–silty admixture. Calcarenite and calcilutite interlayers containing foraminifers locally occur among mudstones. Interlayers of radiolarian cherts are less frequent, and radiolarian assemblages are impoverished as compared to those from Member 1. An increase in sedimentation rate and the higher proportion of fine clastic calciturbidites is apparent here and indicates a greater influx of clastic material from the shelf that resulted in impoverishment of radiolarian assemblages, which, nevertheless, can be correlated with the Oxfordian radiolarians of westerly regions of the Mediterranean Sea (Aita, 1987; Baumgartner, 1984; Baumgartner *et al.*, 1995). The depletion of radiolarian assemblages within the continuous siliceous sequence is untypical of the Mesozoic sections in eastern regions of Russia. Nonetheless, the general trend of sedimentation in the basin did not change; it still developed below the CCD under conditions of good aeration and enhanced oxygen potential.

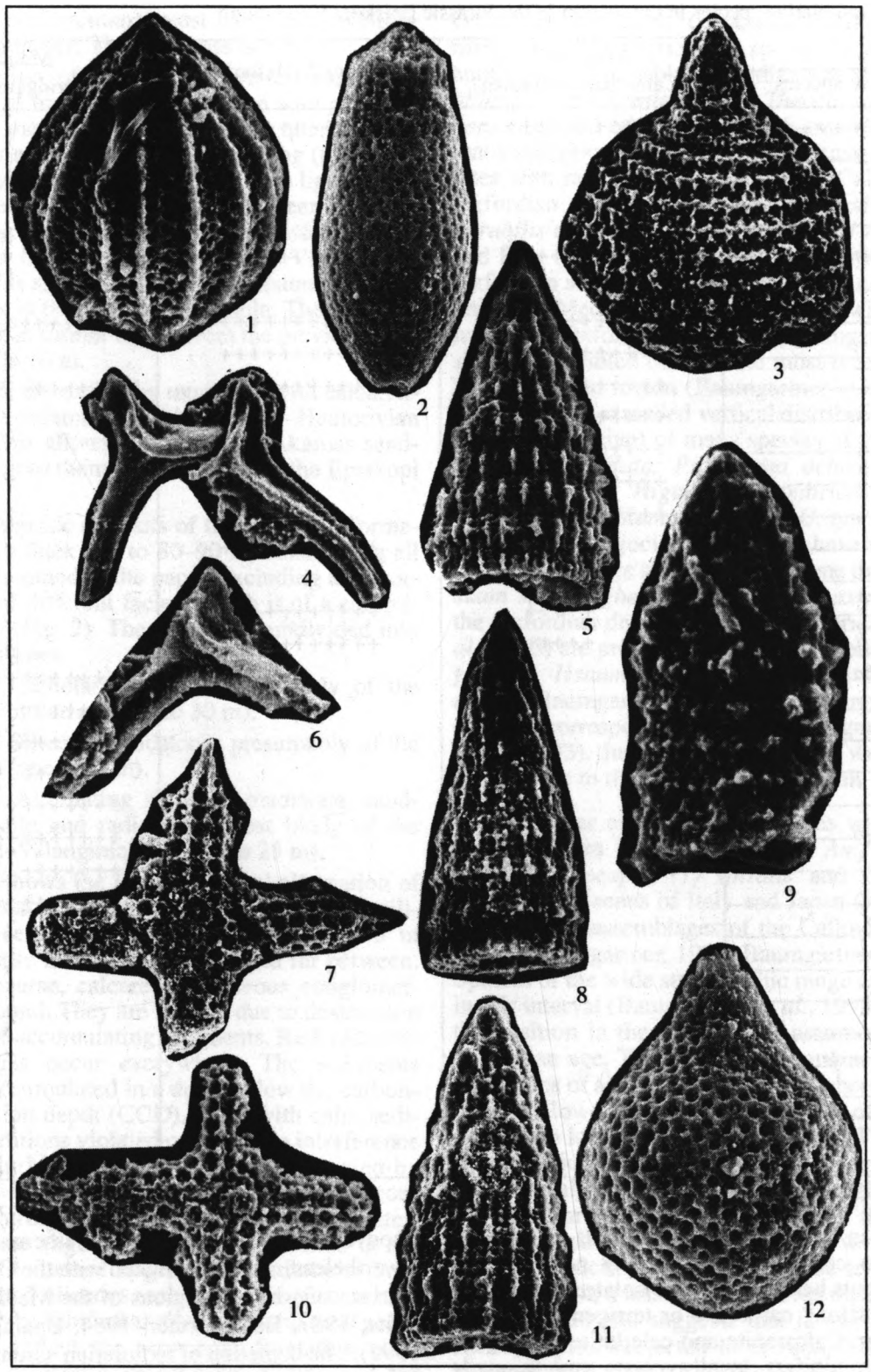

Plate I.

(1) *Protunuma (?) ochiensis* Matsuoka, ×250; (2) *Gueuxella nudata* (Kocher), ×250; (3) *Mirifusus* sp. cf. *M. guadalupensis* Pessagno, ×100; (4) *Acanthocircus suboblongus* (Yao), ×200; (5) *Hsuum maxwelli* Pessagno, ×180; (6) *Acanthocircus* sp. cf. *A. trizonalis* (Rust), ×200; (7, 10) *Higumastra imbricata* (Ozwoldova), both ×180; (8) *Hsuum brevicostatum* (Ozwoldova), ×180; (9) *Parvicingula* sp. cf. *P. dhimenaensis* Baumgartner, ×250; (11) *Semihsuum sourdoughense* Pessagno, Blome et Hull, ×180; (12) *Stichocapsa convexa* Yao, ×250.

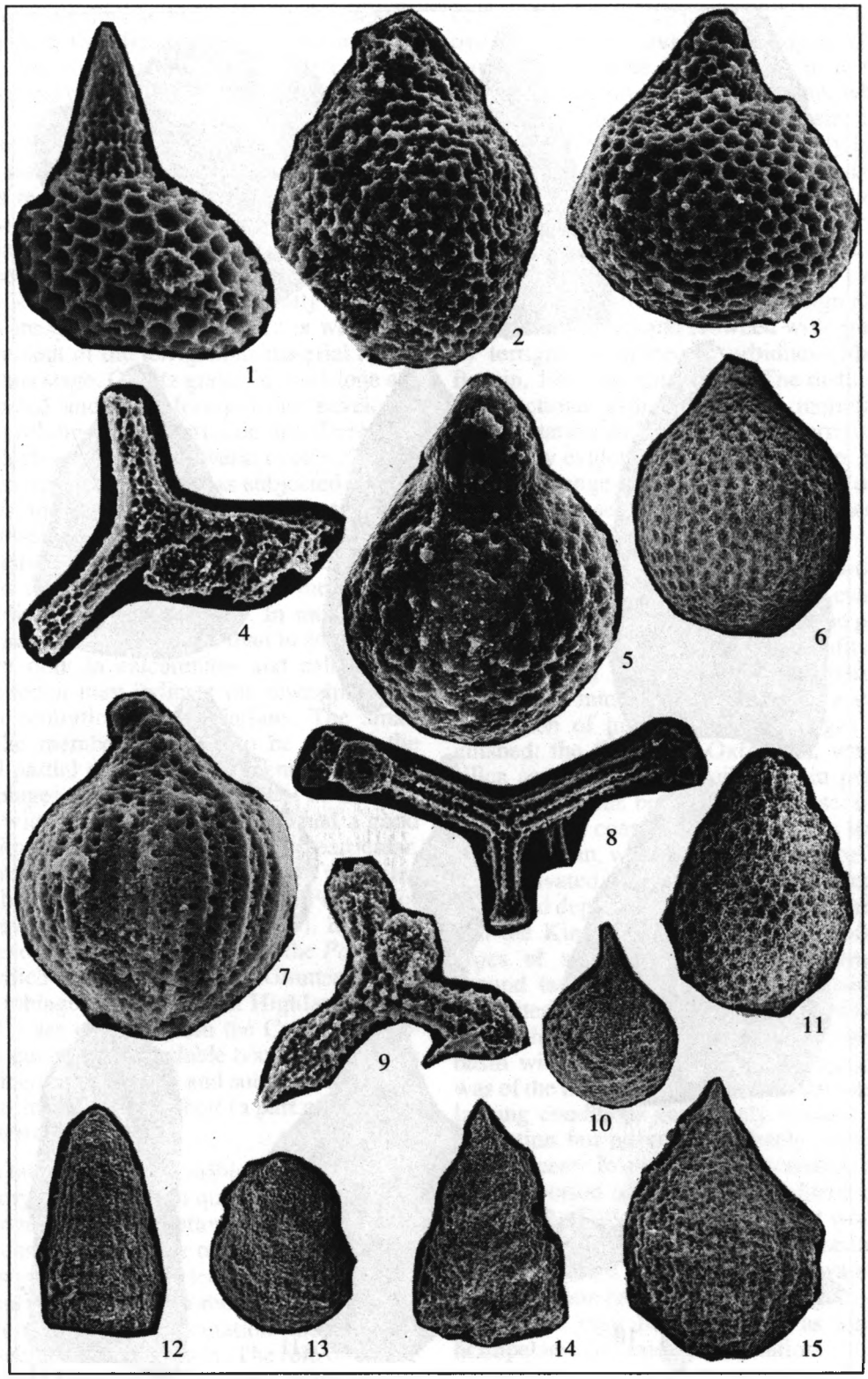

Plate II.

(1) *Stichocapsa* ? sp., ×250; (2, 11) *Stichocapsa himedaruma* Aita, (2) ×250, (11) ×180; (3, 6) *Stichocapsa convexa* Yao, (3) ×250, (6) ×180; (4) *Angulobracchia* sp., ×200; (5) *Stichocapsa* sp. aff. *S. himedaruma* Aita, ×250; (7) *Protunuma* sp. aff. *P. ochiensis* Matsuoka, ×250; (8) *Tritrabs ewingi* (Pessagno), ×180; (9) *Paronaella mulleri* Pessagno, ×200; (10) *Stylocapsa oblongula* (Kocher), ×180; (12) *Archaeodictyomitra* sp. cf. *A. rigida* Pessagno, ×180; (13) *Theocapsomma cordis* Kocher, ×180; (14) *Stichocapsa* ? sp. cf. *S. tairai* Aita, ×180; (15) *Linaresia* sp., ×180.

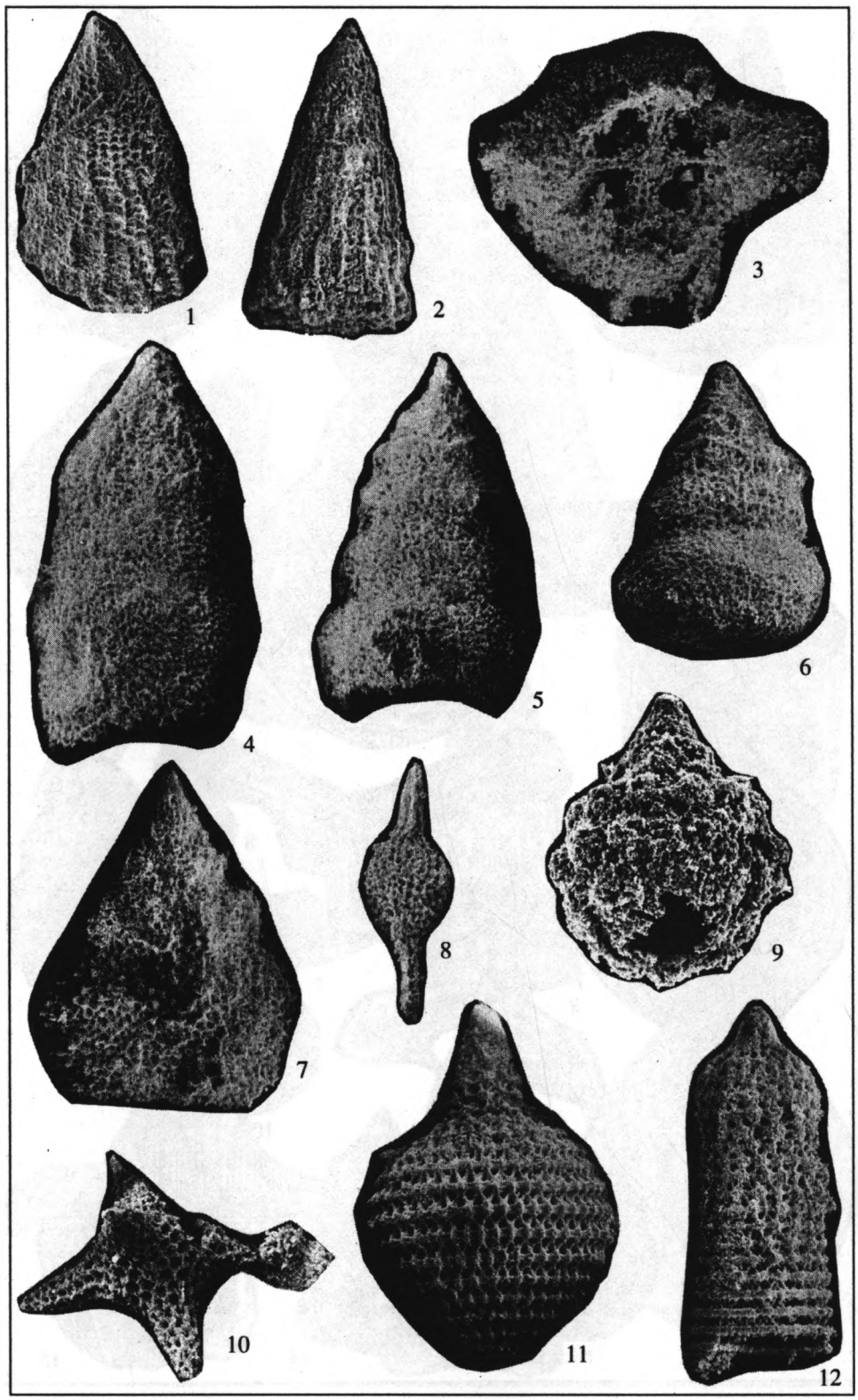

Plate III.

(1) *Semihsuum* sp., ×180; (2) *Hsuum* sp. cf. *H. brevicostatum* (Ozwoldova), ×180; (3) *Higumastra* sp., ×100; (4, 5) *Spongocapsula perampla* Pessagno, both ×100; (6) *Obesacapsula verbana* (Parona), ×75; (7) *Spongocapsula* sp., ×100; (8) *Archaeospongoprunum* sp. cf. *A. patricki* Jud., ×180; (9) *Obesacapsula cetia* Foreman, ×75; (10) *Podocapsa amphitreptera* Foreman, ×100; (11) *Mirifusus dianae* (Karrer), ×100; (12) *Ristola altissima* (Rust), ×180.

Member 3 is most diverse in lithological composition. Key rocks here are siliceous mudstones similar to those from the previous level. The member contains also calcarenites and calcilutites in abundance. Some of them are rich in radiolarians of varying abundance and diversity (Table; Plate III). Radiolarians of the member correlate with assemblages of the Kimmeridgian–Valanginian zones of the West Mediterranean region and Atlantic Ocean (Baumgartner, 1984; Baumgartner *et al.*, 1995). Quartz sandstones distinguished earlier as the Akamas beds (Robertson and Woodcock, 1979; Swarbrick and Robertson, 1980) and slightly silicified marl interbeds are typical of the member as well. The further enhancement of the terrigenous material influx in obvious for this stage. Quartz grains in sandstone are very well rounded and have ferrugineous envelopes indicating the prolonged transportation and derivation from sands, which experienced several cycles of sedimentation and redeposition as well as subjected to aeolian erosion. In the Akamas Peninsula, quartz sandstones are represented by lumps and blocks 10–20 m thick but impersistent along the strike. Structural peculiarities suggest that sandstones represent the channel facies of turbidity and grain currents. In most of the rocks, radiolarians are less frequent than in Member 2, being abundant only in calcarenites and calcilutites. Such a phenomenon may indicate the rewashing and subsequent concentration of radiolarians. The small thickness of the member is likely to be due to the rewashing and partial erosion of sediments. The radiolarian assemblages themselves are typical of an open tropical basin with normal salinity, which had a good connection with other oceanic basins. In particular, they include many species in common with the Kimmeridgian–Tithonian assemblages from the Far East of Russia (Vishnevskaya, 1991; Bragin, 1993). It should be noted, however, that representatives of the *Parvicingula* genus, which are typical of the Kimmeridgian–Tithonian assemblages of the Koryak Highland (Vishnevskaya, 1991) are missing from the Cyprus assemblages. This is quite understandable because the taxon is strictly confined to the boreal and subtropical realms being untypical for the tropical belt (a part of which is the East Mediterranean region).

The data obtained made it possible to infer the Late Mesozoic history of the basin in question. The general regressive changes in sedimentation are apparent from gradually decreasing importance of siliceous rocks and a growing significance of terrigenous deposits. Through the entire section from the base upward, the intensity of terrigenous sedimentation increases to replace the biogenic one completely. The role of coarse clastic turbidity sedimentation becomes more important, whereas the influx of calciturbidites slows down and a progressively greater amount of terrigenous material is transported from the continental block. Only the fine clastic turbidites became the dominating rock types in the Kimmeridgian–Tithonian. Throughout the time interval under consideration, sedimentation occurred either in the marginal part of the basin (Callovian–Oxfordian and Kimmeridgian–Tithonian, with numerous turbidite horizons) or in a small, partially enclosed structure (in the Oxfordian, when radiolarian fauna not typical of the open basin appeared). The region under discussion differs greatly in these features from the Late Paleozoic–Mesozoic basins of northeastern Asia. It has been demonstrated earlier with numerous examples that many Mesozoic sections of siliceous rocks in some regions of the Pacific belt, such as Japan and Sikhote Alin, are extraordinary monotonous in composition (throughout the stratigraphic interval of one–two systems) and crowned with gradual transition by terrigenous–siliceous turbidites (Matsuoka, 1984; Bragin, 1991; Bragin, 1993). The distinct, simple, and unidirectional replacing of the monofacies siliceous sedimentation with deposition of terrigenous turbidites is actually evident in this case. It offers grounds to suggest the change-over from the pelagic (oceanic) to oceanic-margin sedimentation environment (Matsuoka, 1984).

Judging from the Upper Jurassic sections of Cyprus, the history was quite different here. Several interrelated (overprinted) processes operated in the past basin of the region. As the terrigenous non-turbidity (background) sedimentation progressed, it was periodically combined with intensification of silica accumulation and deposition of turbidites. Three stages can be distinguished: the Callovian–Oxfordian, when background silica accumulation prevailed (as in pelagic basins of the Paleo-Pacific but with a higher sedimentation rate) and influx of coarse-clastic turbidites was widespread; the Oxfordian, when the background pelitic sedimentation activated, silica accumulation decreased drastically, and deposition of coarse-clastic turbidites ceased; and the Kimmeridgian–Tithonian, when all the three types of sedimentation (silica accumulation, background terrigenous sedimentation, and formation of turbidites) increased. In the Late Jurassic, this part of the Tethys likely represented the deep-water active basin with the rugged bottom topography. The basin was of the marginal sea or minor (immature) ocean type lacking conditions for strictly oceanic (pelagic) sedimentation but nevertheless freely connected with the World ocean. In general, sedimentation during the considered period of time can be defined as hemipelagic. Only in the Callovian–Oxfordian, it was eupelagic and showed the prevalence of biogenic sedimentation over terrigenous, the good aeration of water, and the low sedimentation rate. The Kimmeridgian–Tithonian sedimentation may be considered as similar to recent hemipelagic oceanic sedimentation.

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The Lower Aptian of the Mangyshlak Mountains

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Abstract—In the Mangyshlak Mountains, the lower Aptian deposits are represented by thin beds (0.5–1.5 m) of sandstone and conglomerate, and in the Kugusem anticline, this time range spans two conglomerate beds incorporating an intermediate clay member (up to 15 m thick). The lower Aptian beds rest with a hiatus on different horizons of the Cretaceous (Berriasian-Barremian), or on the Middle Jurassic rocks. The beds yielded diverse assemblages of various faunal groups: ammonites, belemnites, bivalves, brachiopods, Echinoidea, and others. The beds represent a condensed sedimentary sequence of the shallow-shelf type and are divisible into the lower Aptian *Deshayesites weissi, Deshayesites deshayesi*, and *Dufrenoya furcata* ammonite zones. The lower Aptian ammonite assemblages include 21 species from genera *Deshayesites, Dufrenoya*, and *Cheloniceras*, two of which (*Deshayesites semenovi* and *Cheloniceras sinzowi*) are new and described in the paper. The paper also presents images of 16 other species, including 11 forms discovered for the first time in the Mangyshlak Mountains.

Key words: *Transcaspian region, Mangyshlak, northern Caucasus, southern England, lower Aptian, ammonites, zone, correlation, condensed beds.*

In many Transcaspian areas (Kopetdag, Greater and Lesser Balkhany, Kubadag, and Tuarkyr), the lower Aptian deposits are represented by marine sediments, whose beds are 40 to 650 m thick and span the *Deshayesites tuarkyricus, D. weissi, D. deshayesi*, and *Dufrenoya furcata* zones[1] of the ammonite scale (*Resheniya ...*, 1977; Bogdanova, 1978). Sections of the Mangyshlak Mountains, where fossils of three ammonite zones are encountered in a bed less than 1.5 m in thickness, are unique in the stratigraphic aspect. This bed of massive calcareous sandstone incorporates pebbles and gravel of underlying rocks and phosphorites. It is a basal one in a rather thick sequence of clay with septarian concretions (the Septarium Clay of the Aptian after N.I. Andrusov).

The majority of researchers, who studied the Lower Cretaceous in the Mangyshlak Mountains, believe that the bed of calcareous sandstone or conglomerate (it was frequently referred to as "plate" in publications) spans the total range of the lower Aptian Substage (Luppov, 1932; Savel'ev and Vasilenko, 1963; *Melovye otlozheniya ...*, 1980). On the basis of ammonites represented by index species of different zones, Savel'ev and Vasilenko (1963, p. 271) arrived at the conclusion that the rocks of the "plate" represent the "typical condensed sequence." The lithologic composition of the "plate" was studied by Klycheva and Yakunitskaya (1963), who argued that, in some sections of the Karatau Range, the basal part of the septarium clay above the "plate" may also belong to the lower Aptian.

In 1971 and 1973, Luppov, Bogdanova, and Lobacheva studied 17 exposures of the lower Aptian deposits in the western and eastern Karatau Range, and also in the Karasyaz', Tumgashi, and Kugusem structures (Fig. 1). They outlined the relations of the bed in question with the overlying and underlying deposits, and showed a possibility to subdivide it into the lithologic units in certain localities. Applying in addition the bed-by-bed distribution analysis, Bogdanova identified ammonites and bivalves of the bed, and Lobacheva studied brachiopods[2] and sea-urchins from this unit.

Figs. 2–6 illustrate principal structural features, lithological composition, and distribution of paleontological remains that have been detected in various exposures of the Aptian "plate" within the Karatau and Kugusem structural zones. In all the sections studied, the lower Aptian deposits overlie the uneven erosion surface of underlying deposits ranging in age from the Middle Jurassic (Besokty Formation) to the Hauterivian and Barremian (Kugusem Formation of variegated lagoonal-marine sediments).

In the Chirchili section (near the synonymous well in the westernmost piedmonts of the Karatau Range), the "plate" rests on the dark-gray massive Hauterivian sandstones (*Trigonia* Formation) and represents a gray 0.5-m-thick sandstone bed bearing abundant phosphorite nodules and pebbles variable in size. In addition to

[1] In 1979, Plenum of the Cretaceous Commission of the Interdepartmental Stratigraphic Committee (ISC) added to the lower Aptian the *Turkmeniceras turkmenicum* Zone formerly attributed to the Barremian (*Postanovleniya ...*, 1981; see for my own opinion on the issue at p. 65, point 5).

[2] Brachiopods are described in a separate paper by Lobacheva.

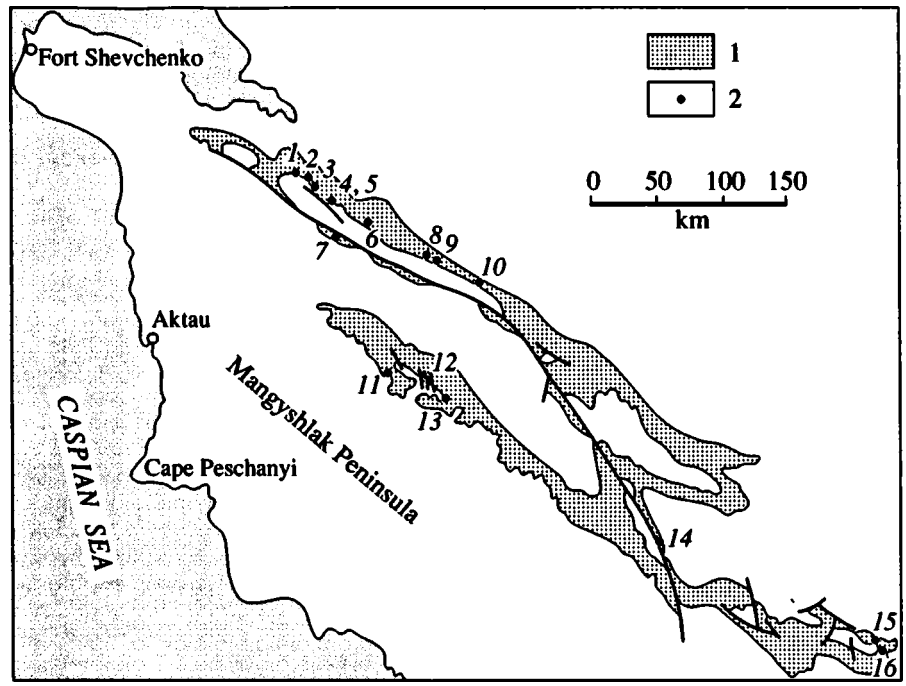


Fig. 1. Geographic scheme of the Mangyshlak Peninsula with outcrops of Lower Cretaceous deposits (1) and the sites studied (2): *1*, Chirchili; *2*, Sorbulak; *3*, Karashimrau; *4*, *5*, Shair; *6*, Kogozbulak; *7*, Tushchibek; *8*, Doshchan; *9*, Dzharmysh; *10*, Kurkruk, *11*, *12*, Karasyaz'; *13*, Sarmurun; *14*, Besokty; *15*, *16*, Kugusem.

ammonites indicated in Fig. 2, the bed yielded plentiful remains of bivalves (*Cucullaea, Lima* s. str., *Grenella, Thetironia, Arctica*, and *Opis*) and brachiopod shells. All fossil remains occur in a form of phosphorite casts or their fragments. The bed is overlain by black smeary clay of the middle Aptian age.

This type of appearance is characteristic of the "plate" under consideration in the Sorbulak, Shair, Tushchibek, Doshchan, Karasyaz', Sarmurun, and Besokty localities.

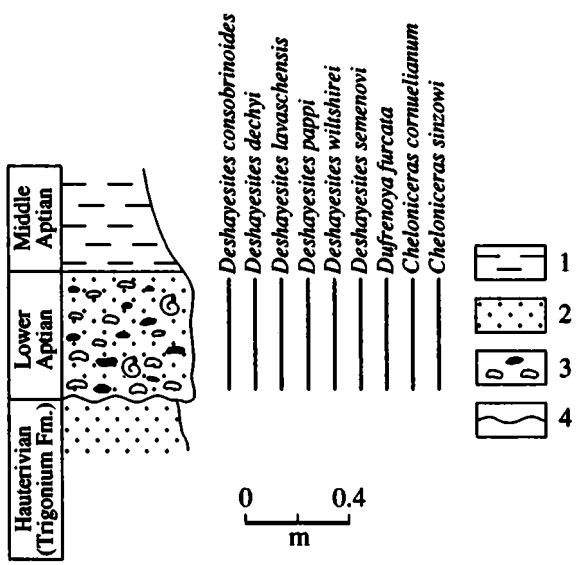


Fig. 2. Chirchili section: (1) clay; (2) sand or sandstone; (3) conglomerate; (4) erosion surface.

In the Karashimrau section (Fig. 3) (7 km to the east from the synonymous wells, the western Karatau Range), the "plate" consists of two beds and overlies the light-green silty clay of the Kugusem Formation. The lower bed of fine-grained rocks incorporates pebbles in its basal horizon only. The calcareous sandstone of the upper bed is intercalated with frequent gravel and conglomerate lentils. Both beds yielded well-preserved fossils. Ammonites of the upper zone of the lower Aptian are recovered from the upper bed only. Both beds are correlative in age with the conglomerate bed of the Chirchili locality.

In the Kogozbulak section (western wall of the Kogozbulak Canyon in the western Karatau Ridge), the "plate" is divisible by lithological features into four beds (Fig. 4). The lower bed of conglomerate with well-rounded pebbles of phosphorite and glauconite rocks is barren of fossils and overlies the erosion surface of grayish green Hauterivian shales. The second bed yielded rare fragments of degesitid shells. Beds 1 and 2 seem to be accumulated earlier than the "plate" rocks in the Chirchili and Karashimrau sections. Beds 3 and 4 include the mixed fauna of all lower Aptian zones and correspond in their formation time to the entire Karashimrau section. Sandstone of Bed 4 grades upward first into its silty variety, and then into the dark, almost black foliate clay enclosing an interlayer with septarium con-

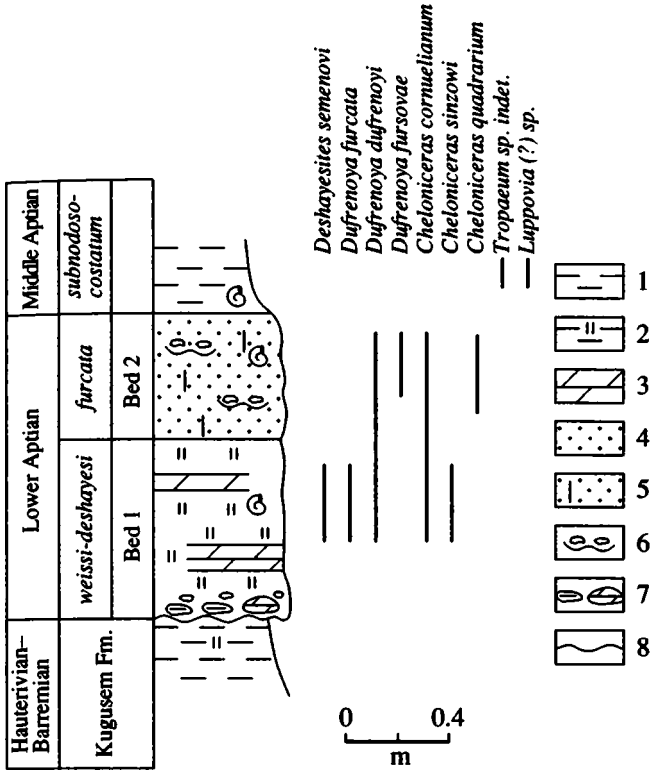


Fig. 3. Karashimrau section: (1) clay; (2) silty clay; (3) marl; (4) sand or sandstone; (5) calcareous sandstone; (6) conglomerate; (7) pebbles of clay and marl; (8) erosion surface.

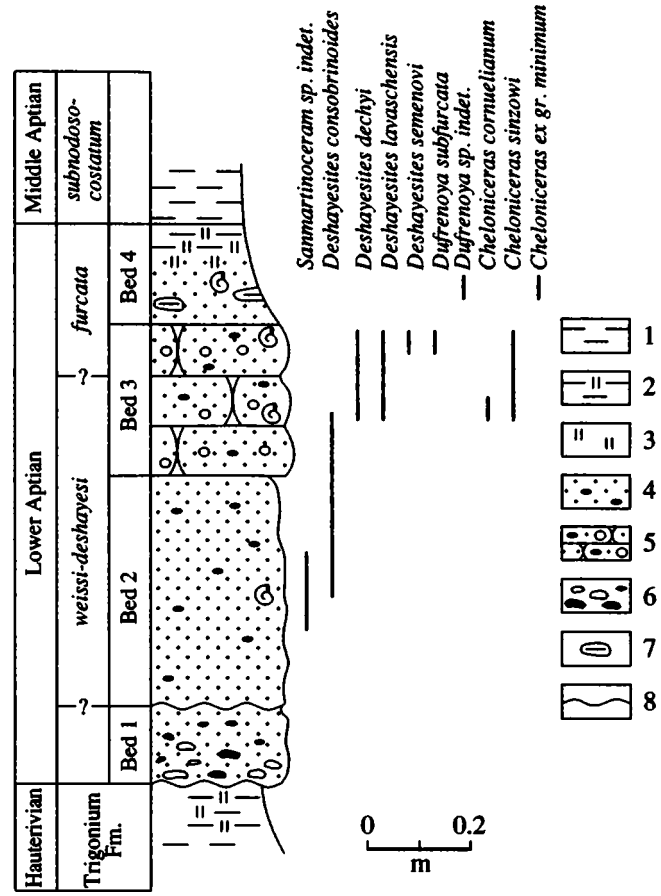


Fig. 4. Kogozbulak section: (1) clay; (2) clayey siltstone; (3) siltstone; (4) sandstone with phosphorite grains; (5) massive oolitic sandstone with phosphorite grains; (6) conglomerate; (7) pebbles of clay rocks; (8) erosion surface.

cretions at the top. The clay yielded remains of the middle Aptian ammonite *Epicheloniceras tschernyschewi* Sinz.

In this section, the lower–middle Aptian boundary does not coincide with the "plate" top and is located above it in the nonconsolidated clay, because the sandstone Bed 4 overlying the "plate" proper bears still the early Aptian ammonites. Since the middle Aptian *Epicheloniceras* forms always occur in Mangyshlak only in the septarium clay and never below it. I place the lower–middle Aptian boundary at the level of 0.5 m above the "plate" top, where the clay is already free of silty or sandy admixture.

In the Dishchan–Dzharmysh section situated between the Doshchan Cemetery in the west and the village of Dzharmysh in the east (the eastern Karatau Ridge), the upper part of the Kugusem Formation (up to 7 m thick) is composed of grayish yellow sandstones cross-bedded at the base and incorporating large concretions (up to 2 m in diameter) of massive calcareous sandstone (Fig. 5). Lentils of conglomerate and gravelstone with frequent phosphorite grains occur everywhere in this member. At the top of the member, there is a bed 1 m thick, and precisely this part of the member is usually considered as a stratigraphic analogue of the Aptian "plate." The bed is composed of yellowish gray, locally bleached sandstone incorporating pebbles and gravel at the base. Only this bed yielded the early Aptian ammonites associated with poorly preserved bivalves of genera *Glycymeris, Aucellina, Chlamys, Camptonectes, Lima* s. str., *Exogyra, Thetironia*, and *Opis*, and also with remains of brachiopods and sea-urchins. In the Doshchan Cemetery area 5–6 km to the west from this section, it is evident that the described bed discordantly overlies the various members of the Kugusem Formation. In this case, the "plate" has a sharp contact with overlying black clay.

In the Kugusem section (Fig. 6) studied in a ravine extending westward from the Kugusem Well, the upper beds of the Kugusem Formation (about 17 m thick in total) are composed, like in the Doshchan–Dzharmysh locality, of sandstones displaying frequently the cross-bedding and enclosing lenses and interlayers of fine-pebble conglomerate and gravelstone. By analogy with many sections of the Karatau Ridge and with the Dosh-

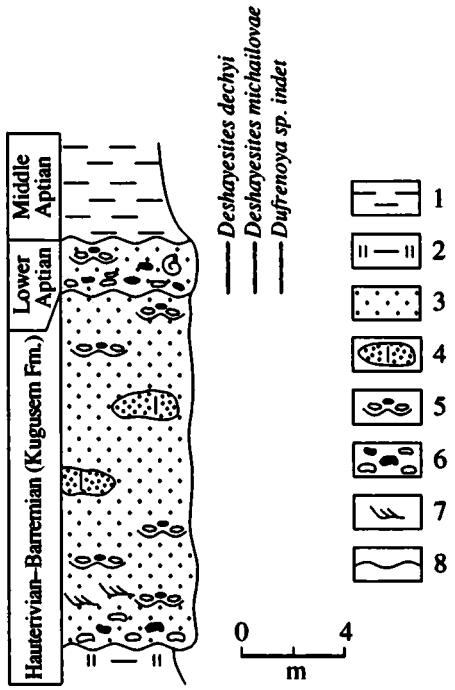


Fig. 5. Doshchan section: (1) clay; (2) clayey siltstone; (3) sand or sandstone; (4) nodules of calcareous sandstone; (5) gravel and conglomerate lentils; (6) conglomerate; (7) cross-bedding; (8) erosion surface.

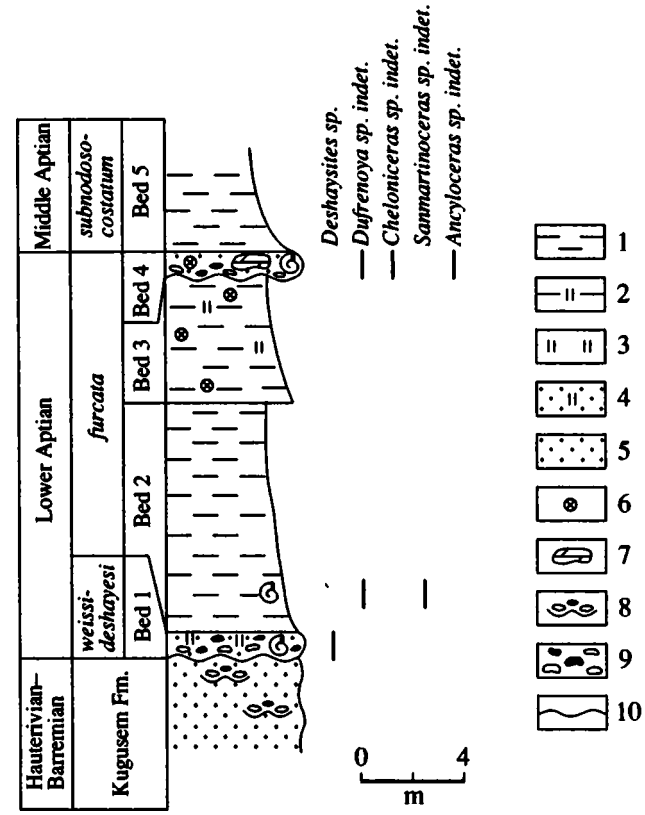


Fig. 6. Kugusem section: (1) clay; (2) silty clay; (3) siltstone; (4) silty sandstone; (5) sand or sandstone; (6) septarium nodules; (7) marl nodules; (8) gravel and conglomerate lentils; (9) conglomerate; (10) erosion surface.

chan–Dzharmysh locality, in particular, the conglomerate Bed 1 at the top of this sandstone sequence is accepted for the basal lower Aptian unit. From this bed, Klycheva and Yakunitskaya (1963) recovered ammonites of the genus *Deshayesites*. In contrast to many other sites, however, the bed does not exhaust here the total range of the lower Aptian. The clay Bed 2 above it yielded the early Aptian *Dufrenoya* forms (Plate II, images 12 and 13; Plate III, image 7). Above the Bed 3 of clayey siltstone barren of fossils, there is another conglomerate bed similar in composition and structure to the lower one and bearing casts of *Dufrenoya* and *Cheloniceras* species. Accordingly, this bed should also be considered as a constituent of the "plate" looking like it is split in two parts by the clay and siltstone interlayers.

By analogy with other sections, the lower–middle Aptian boundary is placed here at the base of the clay Bed 5. This decision is still arbitrary, because the middle Aptian ammonites are extremely rare in the Kugusem area and completely missing from the section in question.

The presented brief description shows that the Aptian "plate" and units below and above it are variable in structure. In the majority of localities however, the lower Aptian interval of the section is represented by either the sandstone, or most frequently, by conglomerate bed up to 1.5 m in thickness, and only in the Kugusem area, this interval includes the 15-m-thick clay unit occurring between two conglomerate beds.

As is already mentioned, Klycheva and Yakunitskaya (1963) additionally attributed to the lower Aptian the basal beds of the septarium clay overlying the "plate." Amid three localities they described, the Chirchili section situated in the southern slope of the eastern Karatau Range is most intriguing. In this section, Klycheva and Yakunitskaya attributed to the lower Aptian the 76-m-thick unit of black clay with remains of *Cheloniceras* sp., *Deshayesites* sp., and *Nuculana pseudomariae* J. Nik., but they did not indicate the position of ammonite remains in the unit. In two other cases (the Keriz and Karakuduk sections), the clay units were dated on the basis of bivalve remains, and such a situation is insufficient to substantiate the early Aptian age of the enclosing rocks. Savel'ev and Vasilenko (1963) also argued that this age suggested for a thick sequence of septarium clay is doubtful and inadequately grounded from the paleontological viewpoint.

The ammonite assemblage from the Mangyshlak Mountains, which was studied in this work, includes 21 species of genera *Deshayesites, Dufrenoya*, and *Cheloniceras*. They occur together with shell fragments of

Table 1. Distribution of ammonites in the lower Aptian section of the Mangyshlak Mountains and other regions

Ammonite species (species marked with asterisk are pictured in Plates I–III)	Mangyshlak Mountains											Turkmenistan						Northern Caucasus					Southern England				
	1	2	3	4	5	6	7	8	9	10	11	1	2	3	4	5	6	1	2	3	4	5	1	2	3	4	5
*Deshayesites babaschensis**					7										+												
*D. consobrinoides**	1			4	15									+	+										+		
*D. dechyi**	17			7	19			4		3	2			+	+					+							
*D. euglyphus**				2									+	+										+			
*D. kudrjavzevi**				7										+	+					+							
*D. lavaschensis**	20		1	16	54															+							
D. levigatus			2											+													
*D. luppovi**					1					1	1		+	+													
*D. michailovae**				1										+	+												
*D. pappi**	2													+													
D. robustocostatus				1																+							
D. semenovi sp. nov.*	1		3	13	29	2																					
D. terminalis		1													+												
D. cf. *wiltshirei**	8	1									1														+		
*Dufrenoya furcata**	3		2	4							1					+					+					+	
*Dufrenoya dufrenoyi**	9		4	10												+											
*D. subfurcata**				15	2											+					+						
*D. fursovae**			1													+											
*Cheloniceras cornuelianum**	10		23	5	13		4	1						+	+	+				+							
*Ch. quadrarium**			1								1														+	+	
Ch. sinzowi sp. nov.*	15		4	10	7			3	3																		
Sanmartinoceras sp.					1																						
Tropaeum sp. indet.					1																						

Mangyshlak sections: (1) Chirchili; (2) Sorbulak; (3) Karashimrau (western); (4) Shair; (5) Kogozbulak; (6) Airakty; (7) Karashimrau (eastern); (8) Tushchibek; (9) Karaduan; (10) Ondy; (11) Karasyaz'; figures in each column indicate amount of collected specimens. Ammonite zones in Turkmenistan (Bogdanova, 1978): (1) *Turkmeniceras turkmenicum*, upper Barremian; (2) *Deshayesites tuarkyricus*; (3) *D. weissi*; (4) *D. deshayesi*; (5) *Dufrenoya furcata*, lower Aptian zones 2–5; (6) *Epicheloniceras subnodosocostatum*, middle Aptian. Ammonite zones of northern Caucasus (Drushchits and Mikhailova, 1966): (1) *Heteroceras astierianum–Imerites giraudi–Colchidites securiformis*, upper Barremian; (2) *Deshayesites weissi-Procheloniceras albrechtiaustriae*; (3) *D. dechyi-D. deshayesi*; (4) *Dufrenoya furcata–D. subfurcata*, lower Aptian zones 2–4; (5) *Cheloniceras subnodosocostatum-Colombiceras crassicostatum*, upper Aptian. Ammonite zones of southern England (Casey, 1961): (1) *Prodeshayesites fissicostatus*; (2) *Deshayesites forbesi*; (3) *D. deshayesi*; (4) *Tropaeum bower* banki, lower Aptian zones 1–4; (5) Cheloniceras martinoides, upper Aptian.

Table 2. Correlated lower Aptian ammonite zonations

Stage, substage	England (Casey, 1961)		Northern Caucasus (Drushchits *et al.*, 1986)	Mangyshlak	Turkmenistan (Bogdanova, 1978)
Lower Aptian	*Tropaeum bowerbanki*	*Cheloniceras meyendorfi*	*Dufrenoya furcata*	*Dufrenoya furcata*	*Dufrenoya furcata*
		Dufrenoya transitoria			
	Deshayesites deshayesi	*Deshayesites grandis*	*Deshayesites deshayesi*	*Deshayesites deshayesi*	*Deshayesites deshayesi*
		Cheloniceras parinodum			
	Deshayesites forbesi	*Deshayesites callidiscus*	*Deshayesites weissi – Procheloniceras albrechti–austriae*	*Deshayesites weissi*	*Deshayesites weissi*
		Deshayesites kiliani			
		Deshayesites fittoni			
	Prodeshayesites fissiocostatus	*Prodeshayesites obsoletus*		Kugusem Formation	*Deshayesites tuarkyricus*
		Prodeshayesites bodei			
Barremian	Weld		*Turkmeniceras turkmenicum–Matheronites ridzewsky*		*Turkmeniceras turkmenicum*

Tropaeum, *Toxoceratoides*, and *Ancyloceras* forms undeterminable at the species level. In many sections of the Aptian "plate" of the Karatau Range, the distribution of zonal index species is indistinct. Nonetheless, when the "plate" is divisible into several beds, one may observe the following distribution patterns: the basal conglomerate is always barren of ammonites (Karashimrau and Kogozbulak localities); the bed above it includes rare degesitid forms (Kogozbulak locality, Bed 2); genera *Deshayesites* and *Dufrenoya* coexist in the middle bed, and the uppermost bed bears the assemblage of *Dufrenoya* and *Cheloniceras* species (Kogozbulak locality, Bed 4; Kugusem locality, beds 2 and 4).

The studied ammonite assemblage and lower Aptian faunas from Turkmenistan (Bogdanova, 1979, 1991) include many species in common. The lower *Deshayesites tuarkyricus* Zone of the Mangyshlak plate yielded both *Deshayesites luppovi* and *D. euglyphus*, whose stratigraphic range in Turkmenistan is wider than this zone. *D. oglanlensis* and index species *D. tuarkyricus*, occurring within the zone only, have not been encountered in the Mangyshlak Mountains, and it is unclear whether this lower Aptian zone is present here or not. It appears that the onset of marine sedimentation in the Mangyshlak region was after the earliest Aptian time. The *Deshayesites wiltshirei* species characteristic of the *grandis* Subzone of the *D. deshayesi* Zone in southern England (Casey, 1964) is established to occur also in Mangyshlak. This fact is of particular interest, because in the Transcaspian areas located southward (Tuarkyr, Greater Balkhan, Kubadag) this part of the Aptian sequence is usually eroded. Other ammonites from Mangyshlak, except local species *Deshayesites semenovi* and *Cheloniceras sinzowi*, are well-known and widespread in many distribution areas of the lower Aptian deposits.

At present, there are two schemes of the lower Aptian ammonite zonation in the Boreal and Tethyan provinces (Table 2). The scheme elaborated by Casey (1961) for southern England exemplifies the boreal zonations, whereas the scheme established in Turkmenistan is accepted as a standard of ammonite zonations in Tethyan regions (Bogdanova, 1978; Hoedemaeker *et al.*, 1993). The lower Aptian ammonite zonation suggested for Mangyshlak (Table 2) is very tentative, because the proper zonal assemblages have not been identified here. Many species available in the Mangyshlak "plate" allow me to believe that they characterize here the zonal units of Turkmenistan and northern Caucasus. At the same time, the lack of the majority of species typical of southern England exclude any chance to apply zonations of the boreal type in the study region. Only two common species, *D. wiltshirei* and *Ch. quadrarium*, both missing from the Transcaspian sections probably because of erosion, characterize the *deshayesi* Zone in both the Mangyshlak region and southern England. The ammonite assemblages of the *furcata* time were very different in the Mangyshlak and England basins. The heteromorphic ammonites, for instance, the genus *Tropaeum*, were dominant at that time in the latter, whereas their abundance in the Mangyshlak and Transcaspian areas is extremely low. The *Tropaeum* remains from basal beds of the septarium clay represent most likely the middle Aptian species of the genus, since they coexist with ammonites of the genus *Luppovia* characteristic of this stratigraphic level in Transcaspian areas.

Judging from the structure of the Aptian "plate" and its faunal remains, there was the following succession of events responsible for the "plate" formation. After the accumulation of the Kugusem Formation in the Hauterivian–Barremian freshwater basin that periodically experienced salinization (Klycheva and

Yakunitskaya, 1963), there was a commencement of the early Aptian transgression. The sea of that time surrounded islands of the flooded Karatau Range. Sediments of the "plate" accumulated in marginal parts of that sea basin and represented a kind of shallow shelf condensation under conditions of frequently alternating sedimentation and erosion phases.

Deposits similar to the lower Aptian "plate" and composed of thin sandstone and conglomerate beds with phosphorite gravel and pebbles are described in many publications, where they are referred to as the condensed beds. For instance, Sapozhnikov and Tashliev (1969) reported about the phosphorite bed of the Gyaurs-Dag locality (Turkmenistan) that yielded fauna of seven Albian zones. They interpreted this bed as a sedimentary unit accumulated after a deep washout of older deposits and under conditions of repeated erosion and sedimentation episodes. In their opinion, the sequence of events led to phosphatization of faunal remains and to formation of pebbles consisting of several phosphate generations. In shoal settings with predominant sedimentation phases, the phosphate beds could be intercalated with sediments of another type, and the same situation seems to be characteristic of the Mangyshlak sections. For example, the described "plate" is locally represented by a thin bed (up to 0.5 m) of phosphorite sandstone bearing mixed fossils of the lower Aptian (Chirchili locality), or the lower and lowermost middle Aptian zones (Besokty locality). In other areas (Kogozbulak locality), the accumulation of sedimentary material above the basal conglomerate dominated over its erosion, as it is evident from the greater thickness of the "plate" (up 1.5 m) and from concentration of older degesitids (genus *Deshayesites*) in its lower part (Bed 2) and younger ammonites in the upper Bed 4. Finally, the condensation and erosion in the Kugusem area took place at the time of the *weissi* and *deshayesi* zones of the lower Aptian, whereas the *furcata* Zone corresponded in time to a period of normal marine sedimentation (clay) in shallow and quiet settings of the past basin. The sequence of events ended with an erosion phase at the early–middle Aptian boundary time. As a result, the condensed conglomerate beds turned out to be split by the intercalated clay.

The suggested model of the lower Aptian "plate" formation certainly needs a verification on the basis of additional lithological and sedimentological data. Nonetheless, the presented data on ammonite distribution in this lithologic unit elucidate to some extent the early Aptian geological history of the region.

The lower Aptian sections similar to those of Mangyshlak are also studied in nearby areas of the northern Caucasus and Dagestan (Rengarten, 1961; Mordvilko, 1962). In Rengarten's opinion, the commencement of transgression induced the strong currents, which washed out the sedimentary material and load off the organic remains, the ammonite shells in particular, thus mixing faunas of different ages. According to data of Rengarten and Mordvilko, the condensed beds of the northern Caucasus include fossil assemblages of two (Tsmur-Chai and Gergebil localities) or three (Levashi and Andiiskoe Koisu localities) lower Aptian zones.[1] Later on, Kakabadze *et al.*, (1978), who additionally investigated the localities in Dagestan, reported that the "load" bed of the Levashi section yielded ammonites of the *dechyi–deshayesi* and *furcata* zones of the lower Aptian mixed with the fossils of the *subnodosocostatum–crassicostatum* Zone of the middle Aptian. The combined Barremian and Aptian faunas were recovered by Sharikadze (1974) near the village of Tsipa from a single conglomerate bed of the Molity syncline, however it is unclear whether this bed is of the condensed type.

The ammonite fauna studied in this work indicate that the transgression in the Mangyshlak region commenced not from the beginning of the degesitid time, but later in the period of the *weissi* Zone. According to data of Mordvilko (1962), the most remarkable stratigraphic level of "coarse sand facies" and condensed beds in Dagestan also corresponds to the upper degesitid zones, and the late Aptian geological history of both regions apparently was similar.

Amid the studied ammonites, there are two new species described below. The collection of ammonites no. 12927 is stored at the Chernyshov Museum of the Central Research Institute of Geological Exploration, St. Petersburg. The original and duplicate parts of this collection include all ammonite species mentioned here in Table 1.

Order Ammonitida

Suborder Ancyloceratina

Superfamily Deshayesitaceae

Family Deshayesitidae Stoyanov, 1949

Genus *Deshayesites* Kasansky, 1914

Deshayesites semenovi[2] Bogdanova, sp. nov.

Plate I, images 1–3

Hoplites weissi Semenov, 1899, p. 112, Plate 3, figs. 12a and 12b

Holotype: no. 1/13927, the Chernyshov Museum, St. Petersburg; Mangyshlak, western Karatau Range, Kogozbulak locality; lower Aptian.

Material: 49 specimens at different stages of evolution.

Description. Large compressed shells displaying the moderate to quick convolution rate. The outer side is narrow, convex, with a smooth outer inflection; lateral sides are slightly convex (whorls are most thick in the lower third of their height); umbilical walls are low and steep in the early whorls, but become more gentle by the growth; umbilical inflection is smooth. In general, the whorl sections are rectangular-oval to triangular-oval. Umbilicus is moderately narrow and stepwise.

[1] I quote here stratigraphic ranges of the beds as they are defined in the author's zonation.

[2] The species name is given in honor of V.P. Semenov, the investigator of Mangyshlak.

Plate I

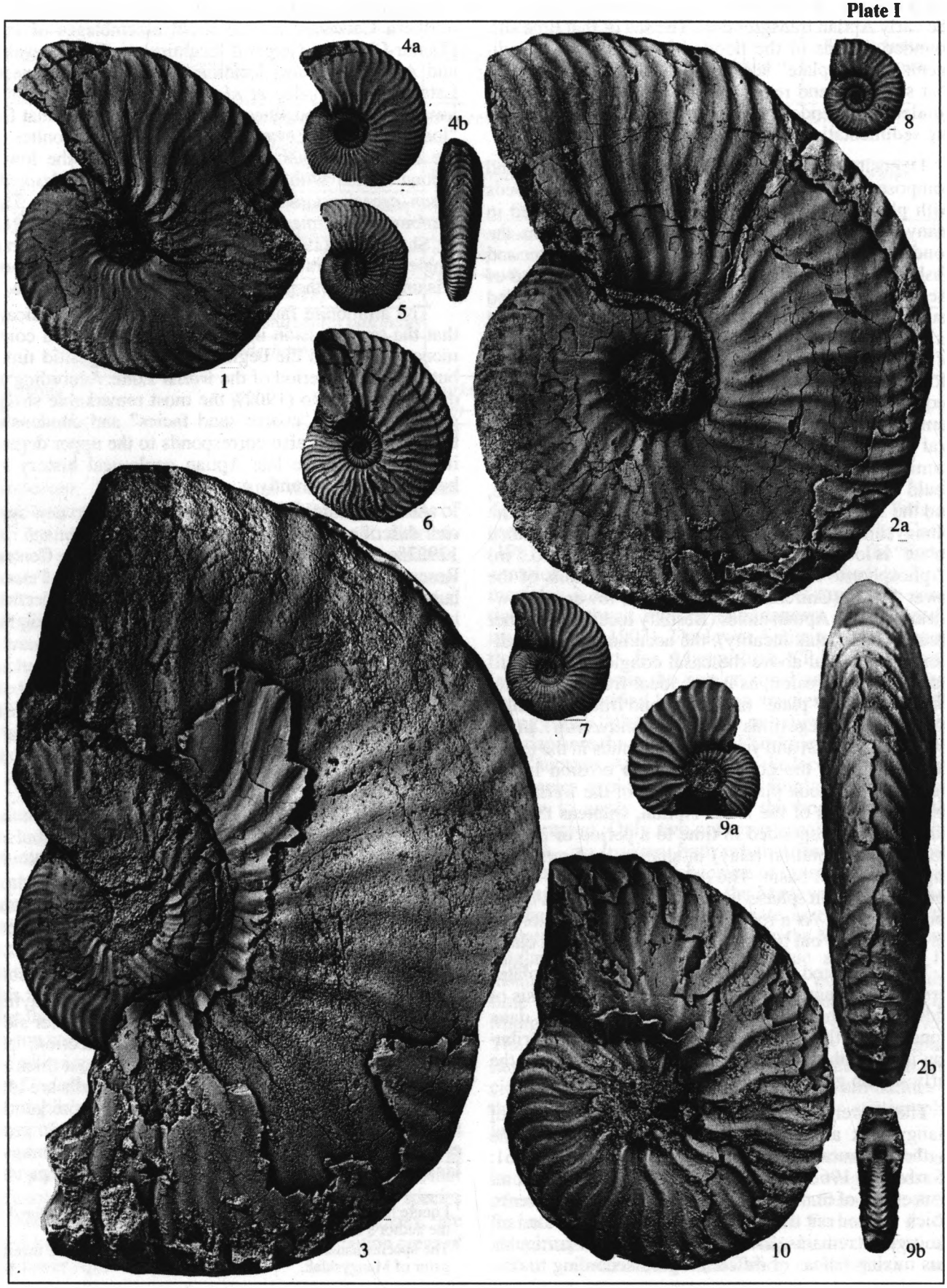


Plate I. Early Aptian ammonites from the Mangyshlak Mountains.

(1–3) *Deshayesites semenovi* Bogdanova, sp. nov.: (1) 2/12927, lateral side; (2) 3/12927, lateral (a) and external (b) sides; (3) holotype 1/12927, lateral side; western Karatau Range, Kogozbulak (talus of beds 3 and 4); lower Aptian. (4–7)*Deshayesites lavaschensis* Kasansky: (4) 10/12927, lateral (a) and external (b) sides, western Karatau Range, Chirchili, lower Aptian; (5) 12/12927, lateral side; (6) 11/12927, lateral side; (7) 13/12927, lateral side, western Karatau Range, Kogozbulak (bed 3); lower Aptian. (8)*Deshayesites pappi* Bogdanova, 11612927, lateral side, western Karatau Range, Chirchili, lower Aptian. (9) *Deshayesites babaschensis* Bogdanova, 14/12927, lateral (a) and apertural (b) sides, western Karatau Range, Kogozbulak (talus of beds 3 and 4), lower Aptian. (10) *Deshayesites luppovi* Bogdanova, 15/12927, lateral side, western Karatau Range, Kogozbulak (talus of beds 3 and 4), lower Aptian.

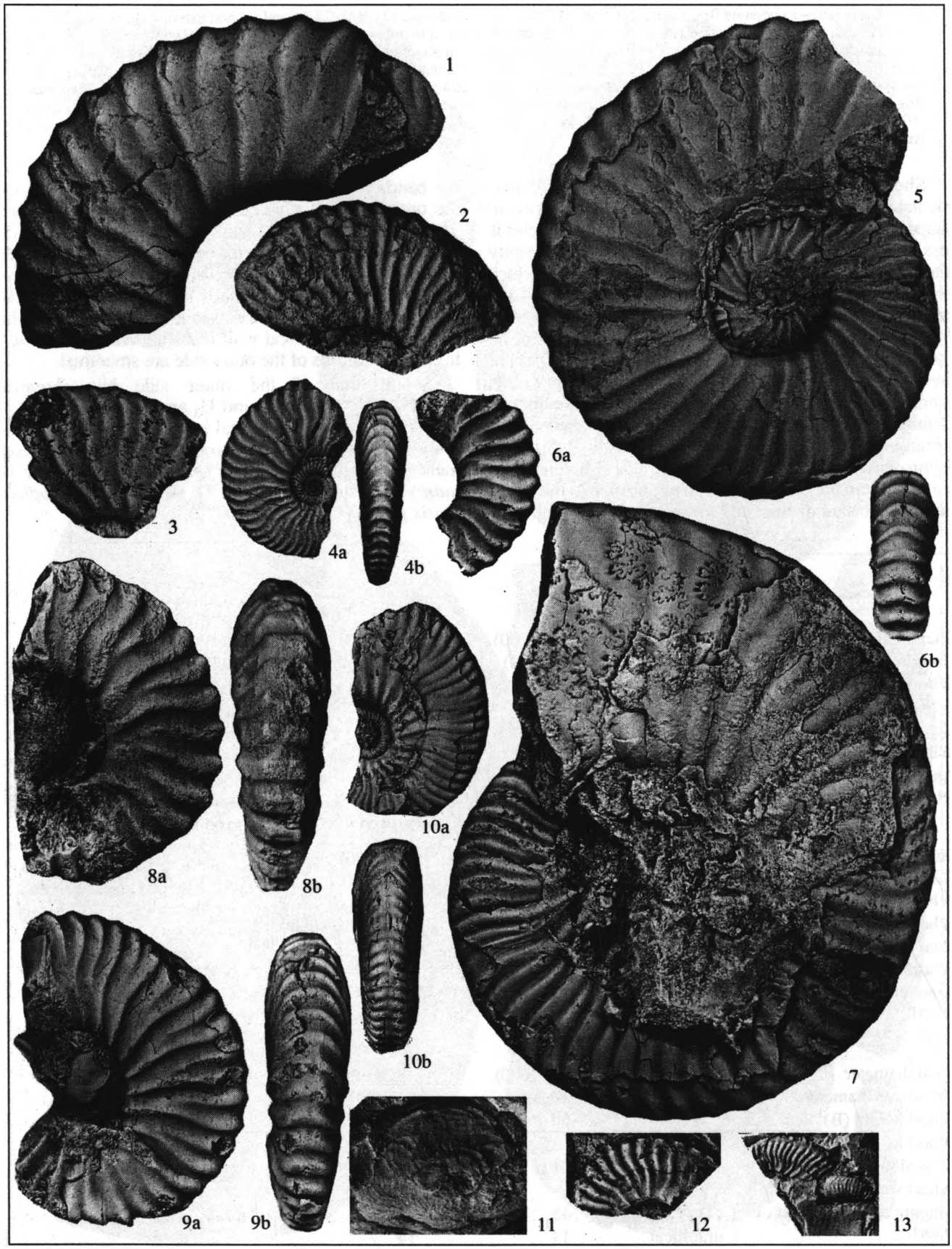
Shells are ornamented with plentiful major and intermediate ribs arranged in a radial manner and changing in shape with the shell growth. When the whorl diameter is less than 30–40 mm, major S-shaped ribs frequently alternate with pairs of intermediate ribs, one of which originates near the umbilical inflection, and another one takes its origin higher, but below the middle of the whorl. Major ribs grow from the basis of umbilical side, extend though the umbilical inflection unchanged in shape, but are less distinct on the outer side of the shell. In whorls greater than 40–70 mm in diameter, the ribs have distinct umbilical crests, and then they become flattened and streaming. Intermediate ribs take their origin at this level (approximately in the middle of the whorl height), and some of them are branching from major ribs. In the upper half, the ribs are distinct and have a shape of flat streaming bands with very narrow intervals between them. On the outer side, all ribs are arc-shaped, convex forward. The mature whorls are usually ornamented with coarse furcate or tripartite major ribs encircling one or two intermediate ribs. All ribs are wide, flattened, and narrow-spaced. Their branching usually begins below the middle of whorls. Origin of major ribs migrates during the evolution from the umbilical wall basis toward the umbilical inflection. All ribs of the outer side are smoothed.

Septal line of the outer side has elements $EIU_1U_3U_4U_{5?}$; lobes U_1 and U_3 are situated on the lateral wall, U_4 at the umbilical inflection, and $U_{5?}$ on the umbilical wall. Judging from the amount and arrangement of umbilical lobes, the species belongs to the *D. tuarkyricus* subgroup of the *D. tuarkyricus–D. oglanlensis* group (Bogdanova, 1979).

Size, mm:

		4/12927	5/12927	6/12927	2/12927
Shell diameter		26.6(100)	29.3(100)	36.2(100)	56.7(100)
Umbilicus diameter		4.7(18)	6.7(22)	7.0(19)	9.2(16)
Whorl height (B)		13.8(52)	14.0(47)	18.5(51)	28.7(50)
Whorl height (b)		6.6	8.5	11.6	18.6
Convolution rate (B/b)		2.1	1.6	1.6	1.5
Whorl width		7.6(29)	8.1(27)	9.6(27)	12.5(27)
Amount of ribs in a half of whorl:	external	29	28	30	26
	umbilical	11	15	10–11	12
		3/12927	7/12927	8/12927	9/12927
Shell diameter		75.8(100)	76.5(100)	80.6(100)	89.2(100)
Umbilicus diameter		16.6(22)	13.0(17)	15.1(19)	19.6(22)
Whorl height (B)		35.2(46)	40.6(53)	38.5(48)	41.6(47)
Whorl height (b)		23.2	24.1	23.5	30.0
Convolution rate (B/b)		1.5	1.7	1.6	1.4
Whorl width		19.0(25)	–	22.0(27)	27.5(31)
Amount of ribs in a half of whorl:	external	32	39	34	33
	umbilical	10	12	10	12
		7/12927	3/12927		
Shell diameter		115.8(100)	118.0(100)		
Umbilicus diameter		20.3(17)	24.0(20)		
Whorl height (B)		59.7(51)	50.3(46)		
Whorl height (b)		38.2	32.1		
Convolution rate (B/b)		1.6	1.6		
Whorl width		–	28.0(25)		
Amount of ribs in a half of whorl:	external	45	32		
	umbilical	14	14		

Plate II



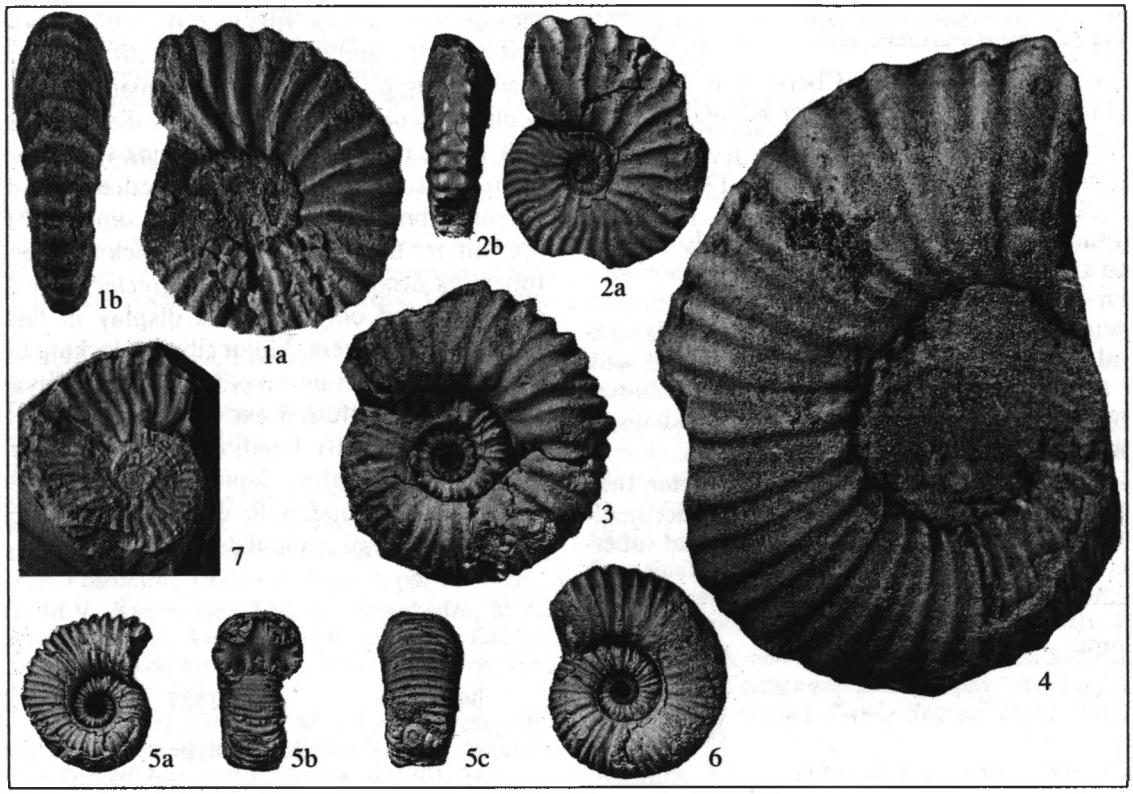


Plate III. Early Aptian ammonites from the Mangyshlak Mountains.

(1) *Dufrenoya furcata* Sowerby, 25/12927, lateral (a) and external (b) sides; western Karatau Range, Shair; lower Aptian (collector N.K. Gordeev). (2) *Dufrenoya dufrenoyi* d'Orbigny, 27/12927, lateral (a) and external (b) sides; western Karatau Range, Shair; lower Aptian. (3) *Cheloniceras cornuelianum* d'Orbigny, 37/12927, lateral side; western Karatau Range, Karashimrau; lower Aptian (Bed 2). (4) *Cheloniceras quadrarium* Casey, 38/12927, lateral side; western Karatau Range, Karashimrau; lower Aptian (Bed 1). (5, 6) *Cheloniceras sinzowi* Bogdanova, sp. nov.: (5) 36/12927, lateral (a), aperture (b), and external (c) sides; eastern Karatau Range, Karaduan, lower Aptian (collected in 1926 by A.P. Il'ina, E.D. Sokolova, and A.G. Eberzin); (6) holotype 32/12927, lateral side; western Karatau Range, Karashimrau; lower Aptian (Bed 2). (7) *Dufrenoya* sp. indet., 30/12927, cast; southeastern Mangyshlak, Kugusem; lower Aptian (clay, *Dufrenoya furcata* Zone).

Plate II. Early Aptian ammonites from the Mangyshlak Mountains.

(1) *Deshayesites consobrinoides* Sinzow, 17/12927, lateral side, western Karatau Range, Shair, lower Aptian. (2) *Deshayesites michailovae* Bogdanova, Kvantaliani, Sharikadze, 18/12927, lateral side, western Karatau Range, Shair, lower Aptian. (3, 4) *Deshayesites dechyi* Papp: (3) 20/12927, lateral side; (4) 19/12927, lateral (a) and external (b) sides, western Karatau Range, Chirchili, lower Aptian. (5) *Deshayesites kudrjavzevi* I. Mikhailova, 21/12927, lateral side (1/2), western Karatau Range, Shair, lower Aptian. (6) *Deshayesites* cf. *wiltshirei* Casey, 23/12927, lateral (a) and external (b) sides, western Karatau Range, Chirchili, lower Aptian. (7) *Deshayesites euglyphus* Casey, 22/12927, lateral side, western Karatau Range, Shair, lower Aptian. (8) *Dufrenoya furcata* Sowerby, 24/12927, lateral (a) and external(b) sides; western Karatau Range, Shair; lower Aptian. (9) *Dufrenoya subfurcata* Kasansky, 26/12927, lateral (a) and external (b) sides; western Karatau Range, Kogozbulak (talus of beds 3 and 4); lower Aptian. (10) *Dufrenoya fursovae* Bogdanova, 28/12927, lateral (a) and external (b) sides; western Karatau Range, Karashimrau (Bed 2); lower Aptian. (11) *Sanmartinoceras* sp. indet., 39/12927, southeastern Mangyshlak; lower Aptian (*Dufrenoya furcata* Zone). (12, 13) *Dufrenoya* sp. indet., shell casts and their fragments: (12) 29/12927, (13) 31/12927, southeastern Mangyshlak, Kugusem, lower Aptian (*Dufrenoya furcata* Zone).

Comparison. The described species differs from many ammonites of the indicated group as having the narrower umbilicus. The lesser amount of ribs and the lack of their fasciculation differ it from *D. weissi* Neum. et Uhl. (Neumayr and Uhlig, 1881, p. 51, Plate 46, figs. 1 and 1a). *D. callidiscus* Casey (Casey, 1964, p. 327, Plate 49, figs. 3 and 4; Plate 51, fig. 3; Plate 53, fig. 2) has another shape of the whorl section and a shorter interval of rib flattening on the lateral side. In comparison with *D. similis* Bogdanova (Bogdanova, 1991, p. 84, Plate 1, figs. 2 and 3; Plate 2, figs. 7 and 8), the new species has a lesser number of ribs lacking fascicles.

Distribution: Mangyshlak Mountains, western Karatau Range, Chirchili, Sorbulak, Karashimrau, Shair, Kogozbulak, and Airakty localities; lower Aptian.

Cheloniceras sinzowi[1] Bogdanova, sp. nov.

Plate II, figs. 5 and 6

[1] The species name is given in honor of well-known paleontologist I.F. Sinzow.

Cheloniceras seminodosum: Sinzow, 1906, p. 165, Plate 1, figs. 3a, 3b, 3c

Holotype: no. 32/12927, the Chernyshov Museum, St. Petersburg; Mangyshlak, western Karatau Range, Karashimrau locality; lower Aptian.

Material: 52 specimens of various preservation degree.

Description. Small shells with extremely convex whorls and a moderate convolution rate. The whorls are rectangular-oval in sections, and their width is greater than the height. The outer side is wide, moderately convex; lateral sides are low, convex, and connected with the outer side through a smooth inflection; umbilical wall is low and steep; umbilical inflection is indistinct. The stepwise umbilicus is wide.

When whorls are less than 25 mm in diameter, they are ornamented with radial ribs (major and intermediate) and with two rows of umbilical and lateral tubercles or crests. Major ribs take their origin at the suture and cross the umbilical inflection displaying the geniculate bending. Major ribs have marginal tubercles or acute geniculate ridges. The less distinct intermediate ribs frequently show the posterior type of branching from the ridges. Isolated intermediate ribs take their origin at the same low level of the umbilical wall; they are thinner than major ribs and lacking tubercles. All tubercles disappear in whorls greater than 25 mm in diameter, and only rare ribs display in this case an indistinct roughness. Major ribs are lacking bifurcation and encircle short intermediate ribs. All ribs are almost identically manifested and cross the outer side under right angle slightly bending forward. In some specimens, ribs are slightly depressed in the middle zone of the outer side, and both ends of the depression are marked by minor geniculate bends.

Size, mm:

		33/12927	34/12927	32/12927 Holotype	32/12927
Shell diameter		26.6(100)	25.9(100)	31.7(100)	36.0(100)
Umbilicus diameter		9.0(34)	–	12.5(40)	12.4(35)
Whorl height (B)		11.0(41)	9.0(3.5)	10.6(30)	13.5(37)
Whorl height (b)		7.4	6.0	8.2	9.9
Convolution rate (B/b)		1.49	1.50	1.30	1.36
Whorl width		14.5(54)	13.8(53)	16.2(51)	20.0(55)
Amount of ribs in a half of whorl:	external	21	–	23	–
	umbilical	13	–	16	–

Comparison. As compared with *Ch. seminodosum* Sinzow (Sinzow, 1906, p. 164, Plate 1, figs. 4–6), the new species has the narrower whorl sections, thinner ribs, and does not display the stage of newly appearing tubercles. The very similar *Ch. minimum* Casey from England also represents a microconch (Casey, 1962, p. 217, Plate 35, figs. 5a–5c) and has tubercles and rib bifurcation in the last whorl clearly seen in fig. 5a of the reference work, but lacking in the described species. The latter has certain similarity with *Ch. cornuelianum pygmaeum* Nikschitsch (Nikschitsch, 1915, p. 17, Plate 1, figs. 7 and 8). Unfortunately, this researcher pictured only an initial convolution stage of this form (up to diameter of 20 mm) inappropriate for the correct comparison.

Distribution: Mangyshlak Mountains, western Karatau Range, Chirchili, Karashimrau, Shair, Kogozbulak, and Tushchibek localities; eastern Karatau Range, Dzhaprakty and Karaduan localities; lower Aptian.

ACKNOWLEDGMENTS

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New Locality of the Campanian Radiolarians in Southwestern Sakhalin

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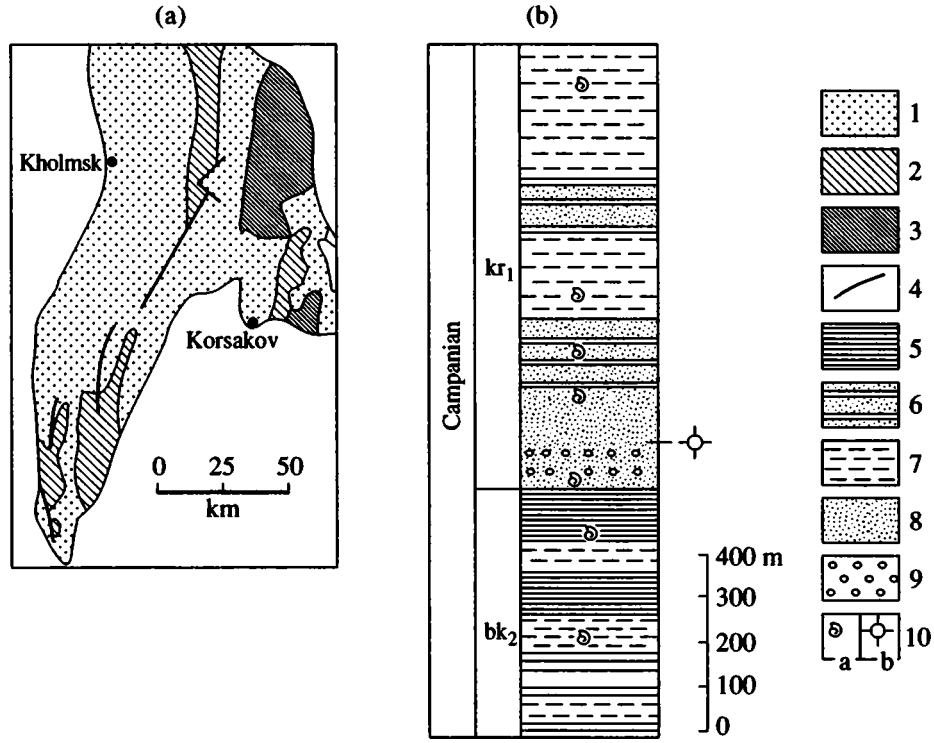
Abstract—The representative radiolarian assemblage different in taxonomy from all known coeval assemblages of Sakhalin was discovered in the basal beds of the Lower Krasnoyarka Subformation of the upper Campanian in the Susuya River basin of southwestern Sakhalin. It appeared to be hardly correlative with coeval assemblages from adjacent and remote areas. The article presents descriptions of the new genus *Bipylomella* gen. nov. and new species *Bipylomella sachalinica* sp. nov., *Hexaloche (?) horridus* sp. nov., and *Spongostaurus (?) sachalinensis* sp. nov.

Key words: *stratigraphy, Sakhalin, Campanian, radiolarians, correlation, endemism.*

In the West Sakhalin mountains, Campanian radiolarians were known from the Krasnoyarka Formation (*Opornyi ...*, 1987) exposed along the Kholmsk highway to the west of the town of Yuzhno-Sakhalinsk. During the field investigation of 1992, a new radiolarian locality was discovered in the Susuya River valley near the Sinegorsk Settlement. The Krasnoyarka Formation section is similar here to the stratotypical one in the Krasnoyarka River basin. It is represented by medium- to fine-grained gray calcareous sandstones intercalated with gray siltstone interbeds, where frequent and big (up to 30–40 cm in diameter) carbonate concretions contain *Schmidticeramus schmidti* (Mich.), *Inoceramus sachalinensis* Sok., and *I. orientalis* Sok. The sandstone beds are overlain by dark-gray siltstones with fine-grained gray sandstone interbeds enclosing carbonate concretions, which bear *Conadoceras* sp., and *Pachydiscus* cf. *neubergicus* Hauer (the figure).

One concretion from the lower part of the sandstone member of the Krasnoyarka Formation (80 m above the base) yielded radiolarians *Crucella espartoensis* Pessagno, *C. zonovae* Kasinzova, *Cuboctostylus kasinzovae* Bragina (Bragina, in press), *Hexalonche (?) horridus* sp. nov., *Patelluta verteroensis* Pessagno, *Patulibracchium* cf. *petroleumensis* Pessagno, *P. vereshagini* Kasinzova, *Phacodiscus* (?) sp., *Phaseliforma carinata* Pessagno, *Bipylomella sachalinica* sp. nov., *Pseudoaulophacus floresensis* Pessagno, *Spongostaurus (?) hokkaidoensis* Taketani, *S. (?) sachalinensis* sp. nov., *Amphipyndax stocki* (Campbell et Clark), *A.* aff. *ellipticus* Nakaseko et Nishimura, *Archaeodictyomitra striata* (Lipman), *A. squinaboli* Pessagno, *Clathrocyclas* sp., *Eucyrtidium* sp. A, *Lithostrobus* cf. *zhamoidai* Kasinzova, *Stichomitra livermorensis* (Campbell et Clark), *Theocampe guttaeformis* Bragina, and *Vistylaria magna* Gorka, and *Stylodruppa bifascicula* Kasinzova.

The assemblage consists of well-preserved diverse radiolarians and shows the close taxonomical similarity to the coeval *Pseudoaulophacus floresensis–Stichomitra livermorensis* assemblage described by Kazintsova (1979; *Atlas ...*, 1993). However, there is a number of species, which were not recorded in the coeval assemblages of the region. Radiolarians are abundant and diverse in morphology; all species are in equal proportions except for some prevailing representatives of the Actinomidae Family. The radiolarians are mostly thin-walled (Plate I, nos. 1, 5, 8; Plate III, nos. 1, 2, 9) (*Stichomitra livermorensis* and *Amphipyndax stocki*). Cellular forms (Actinommidae, Plate II, nos. 1, 2, 4–6) are numerous. Discoidal and prunoidal morphotypes of the assemblage have, as a rule, the spongy shells. Shells have either smooth (*Amphipyndax ellipticus*) or, more frequently, spinous (*Amphipyndax stocki*) surfaces. Some species demonstrate regularly spaced round pores (*Vistylaria magna*) or polygonal pore frames (*Lithostrobus zhamoidai*), while other species are characterized by irregularly spaced pores of variable form and size (*Spongostaurus (?) sachalinensis* sp. nov.). There are some forms with pseudoaulophacoidal wall structure (*Pseudoaulophacus floresensis*). An interesting feature is the heavy patagium cover distinct in representatives of *Patulibraccium petroleumensis*. This feature is characteristic of the same species found at the Kholmsk highway (*Opornyi ...*, 1987, Plate XXXV, Fig. 9). Some Actinommidae forms show very complex structure consisting of several envelopes. In most cases, it is impossible to distinguish the complicatedly branching medular shell. Kazintsova (*Atlas ...*, 1993) reported on a number of species, which have not been described in publications. Most of them are very big Sphaerellaria forms characterized by a complex shell structure and some representatives of genera *Eucyrtid-*



The Upper Cretaceous sequence of the Susuya River area (the West Sakhalin Mountains), schematic geological map (a) and lithological column (b): (1) Cenozoic deposits; (2) Upper Cretaceous deposits; (3) pre-Cretaceous deposits; (4) fault; (5) siltstone; (6) sandstone with siltstone interbeds; (7) clay; (8) sandstone; (9) carbonate concretions; (10) fauna of (a) inocerams and (b) radiolarians; (kr) Krasnoyarka Formation; (bk) Bykov Formation.

ium, Clathrocyclas, and *Phacodiscus*. The analysis revealed some unusual discoids having two pylomas, which had not been earlier detected in the Cretaceous deposits. Another characteristic feature of the assemblage is the presence of spicule-bearing Sphaerellaria forms, namely, of representatives of genus *Cuboctostylus* (*Cubostostylus kasinzovae* Bragina) (Bragina, in press). This find is significant because no spicule-bearing Sphaerellaria have been so far known from the Upper Mesozoic deposits. Only recently, Dumitrica (1994) described some Early and Middle Cretaceous spicule-bearing forms, that he attributed to the new genus *Pyloctostylus*. The occurrence of such forms far from the Mediterranean region and California (the more so, belonging to a different genus found in Sakhalin for the first time; Bragina, 1997) improves our knowledge of the distribution pattern of spicule-bearing forms and allows the inference that these taxa were not relict in the Late Cretaceous. Their scarcity in radiolarian assemblages can also be explained in other way. Being very brittle, these big cellular radiolarians are unsuitable for preservation, and the possibility of finding them depends on lithological features of enclosing rocks. In our case, the carbonate concretions appear to be an ideal material for the brittle cellular forms to be preserved.

In southwestern Sakhalin, Kazintsova (*Atlas ...*, 1993) distinguished two radiolarian assemblages of the Campanian age: (1) the early Campanian *Spongostaurus (?) hokkaidoensis–Hexacontium* sp. assemblage and (2) the late Campanian *Pseudoaulophacus floresensis–Stichomitra livermorensis* assemblage. The assemblage under consideration is of a late Campanian age but it includes *Spongostaurus (?) hokkaidoensis*, which was described from the Campanian deposits of Hokkaido (Taketani, 1982) and taken for the Campanian zonal species of Japan. Therefore, the occurrence of this species in the late Campanian assemblage does not mean that it is the characteristic form of this substage in Sakhalin (*Atlas ...*, 1993). The inference that it ranges throughout the Campanian would be more correct.

The taxonomic analysis shows that the species known in California (Campbell and Clark, 1944; Pessagno, 1976) represent 30% of the assemblage. At the same time, only few species in common (*Amphipyndax stocki, Archaeodictyomitra squinaboli*, and *Pseudoaulophacus floresensis*) are present in the coeval cold-water assemblages of the Middle Volga region (Bragina *et al.*, in press), two other species (*Patulibrachiium petroleumensis* and *Amphipyndax stocki*) are known from the Santonian–early Campanian (?) assemblage of the Moscow region (Bragina, 1994), one species

Amphipyndax stocki is described by Gorbovets as *Dictyomitra uralica* from the coeval assemblage of the West Siberian Lowland (Kozlova and Gorbovets, 1966), and three species (*Crucella espartoensis, A. stocki*, and *Vistularia magna* discovered in the Far East for the first time) are characteristic of Campanian assemblages of Poland (Gorka, 1989).

The regions mentioned above are quite distant from the West Sakhalin Mountains. How similar are these assemblages to those of the Far East? The closest taxonomic similarity is demonstrated by the late Santonian–early Campanian assemblage of the Penzhina Bay (Bragina, 1991b), which contains, in addition to *Theocampe guttaeformis* and *Amphipyndax stocki*, some fragments of *Cubostostylus kasinzovae* and five new undescribed species of the Actinommidae family. It should not be excluded that this similarity between assemblages can be a result of exceptionally favorable burial conditions. The good preservation of radiolarians was controlled by lithological features of the enclosing deposits as in both cases radiolarians are macerated from carbonate concretions. The comparison with another Far East radiolarian assemblage of late Campanian–Maastrichtian age from the Shikotan Island (Bragina, 1991a) revealed their almost absolute difference, because the latter is younger in age.

The assemblage under consideration has only one species in common with the Campanian–Maastrichtian assemblage of the Valaginskii Ridge (Vishnevskaya and Bernard, 1986).

The Campanian radiolarian assemblages of the Shimanto Group (southwestern Japan) consisting of taxa typical of tropical areas (Nakaseko and Nishimura, 1982; Hashimoto and Ishida, 1997) have one species (*Amphipyndax stocki*) in common with the assemblage under discussion. The radiolarian assemblages of Hokkaido (Taketani, 1982) contain three species of this kind: *Amphipyndax stocki, Spongostaurus (?) hokkaidoensis*, and *Crucella espartoensis*. Foraminiferal assemblages of Sakhalin and Hokkaido are also quite different (Taketani, 1982; Maiya and Takayanagi, 1977), whereas there are many features in common (over 50% of taxa) between coeval inoceram assemblages of these regions (*Atlas* ..., 1993; Matsumoto, 1977) The low degree of similarity between radiolarian and foraminiferal assemblages from coeval deposits of the Hokkaido–Sakhalin paleobasin can be explained by inadequate knowledge of microfauna from different localities of the region. Another explanation is that Sakhalin faunal assemblages developed in another paleoclimatic zone should not be excluded as well.

CONCLUSION

This investigation resulted in distinguishing the radiolarian assemblage including some taxa, which have not been recorded earlier in the region. The occurrence of spicule-bearing forms is of particular significance. The new genus *Bipylomella* and new species *Bipylomella sachalinica* sp. nov., *Spongostaurus (?) sachalinensis* sp. nov., and *Hexalonche (?) horridus* sp. nov. are described below. The comparative analysis showed that the distinguished assemblage has little in common with coeval radiolarian assemblages from remote and adjacent regions. Few species identical to those from the Campanian assemblages of the Hokkaido and Shikoku islands may give evidence either for the inadequate study of microfaunas or for paleoclimatic differences. The close similarity between the assemblage in question and remote subtropical assemblages may indicate thorough studies of the latter.

The radiolarian fauna of the Pacific paleogeographic region is considered to be highly endemic. The analyzed composition of new assemblage confirms this opinion. The assemblage includes a number of species to be carefully studied and described.

SYSTEMATIC DESCRIPTIONS

Family Spongodiscidae Haeckel, 1882

Genus *Bipylomella* gen. nov.

Type species: *Bipylomella sachalinica* sp. nov.

Name: *Bipylomella* (Latin.), having two-pylomes.

Description. Large biconvex discoidal shell of a subtriangular form, which has two pylomes oriented toward the apices. The shell is filled up with fine spongy tissue.

Comparison. Representatives of *Bipylomella* gen. nov. differ from those of genus *Spongopyle*. They have two pylomes and show the absence of distinct concentric layers in the internal meshwork.

Remarks. Few discoidal and prunoidal taxa having pylomes were found in Late Cretaceous radiolarian assemblages. This skeletal element is characteristic of Cenozoic and recent forms (Dreyer, 1889; Petrushevskaya, 1967). For instance, the generic identification of the Cretaceous species *Spongopyle insolita* Koslova is based on the presence of one pylome and internal concentric layers. This genus was described by Dreyer (1889) from the recent radiolarian assemblages of the Indian Ocean. In spite of a long period of studies of pylome-bearing taxa, the interpretation of the term "pylome" is still ambiguous. Dreyer (1889), who was the first to use this term, meant the cone-like deepening, which represents a tube connected, according to Petrushevskaya (1967), with radial spines. However, a cone-like hole may also occur in forms with the spongy inner tissue. Being differently interpreted, the term "pylome" cannot be used with confidence for paleontological description. The problem requires thorough investigations of the inner structure of the Cenozoic and recent taxa of this kind.

Bipylomella sachalinica sp. nov.

Plate I, 1–3

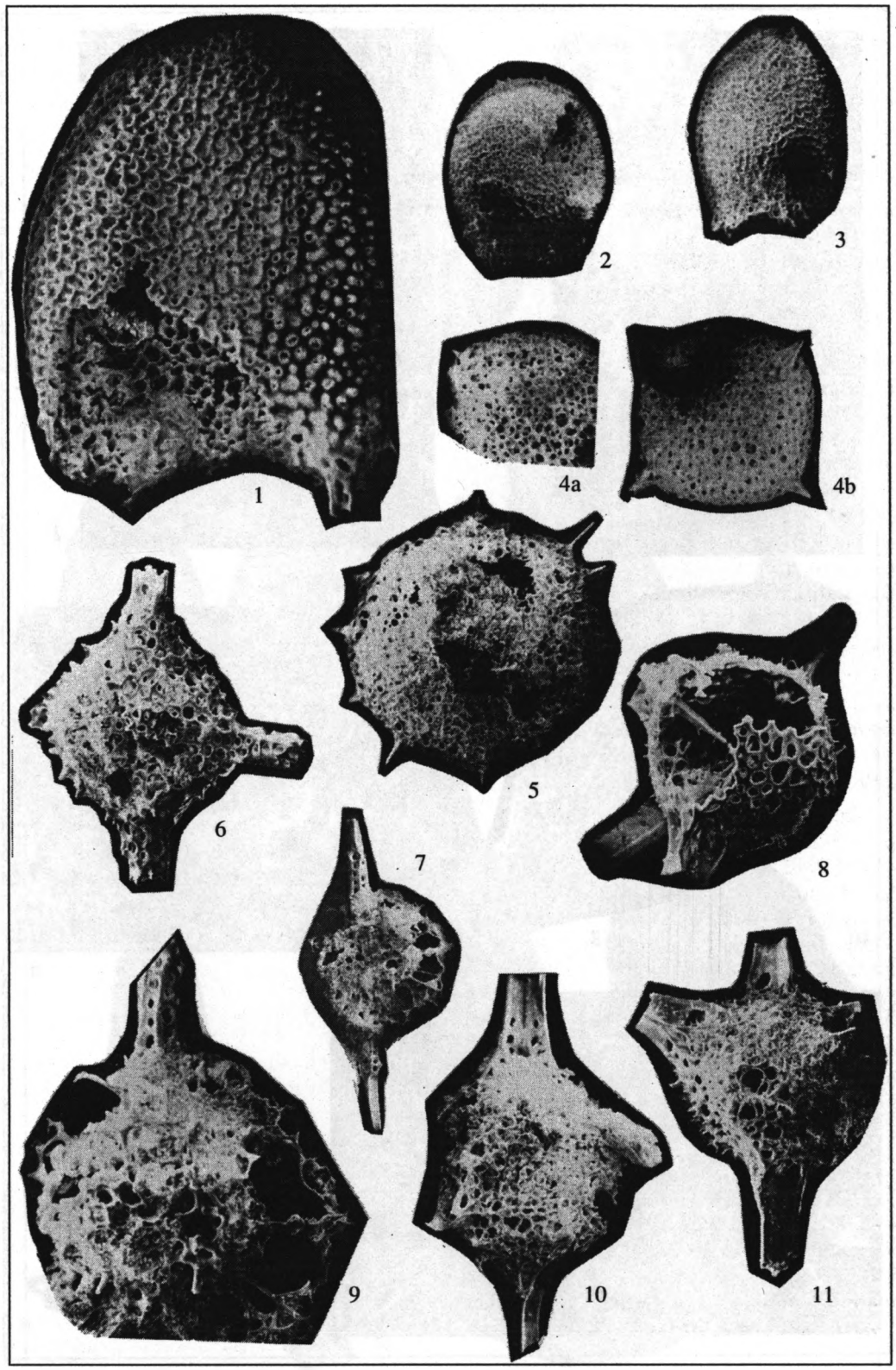


Plate I. Late Campanian radiolarians of southwestern Sakhalin (Plates I–III).
(1–3) *Bipylomella sachalinica* sp. nov., (1) × 500; 2, 3 – × 200. (4a, 4b) *Spongostaurus* (?) *sachalinensis* sp. nov., ×200. (5) *Phacodiscus* sp., ×200. (6, 7) Gen. et sp. indet. A, (6) ×200; (7) ×100. (8) Gen. et sp. indet., ×200. (9) Gen. et sp. indet. A, ×500. (10, 11) Gen. et sp. indet., ×200.

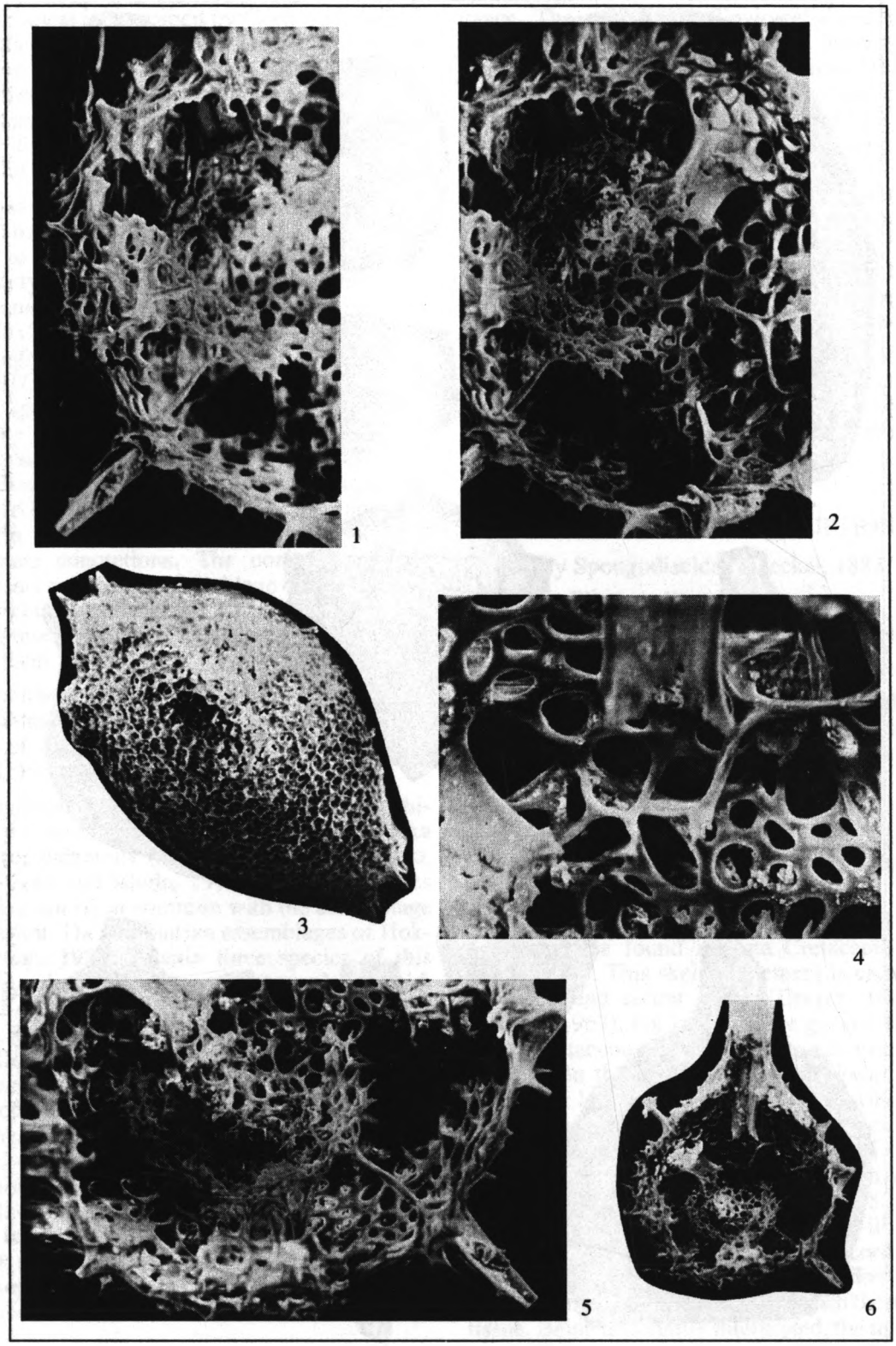


Plate II. (1, 2) *Hexalonche* (?) *horridus* sp. nov., ×500. (3) *Stylodruppa bifascicula* Kasinzova, ×500. (4) *Hexalonche* (?) *horridus* sp. nov., ×1000. (5) *Hexalonche* (?) *horridus* sp. nov., ×500. (6) *Hexalonche* (?) *horridus* sp. nov., ×200.

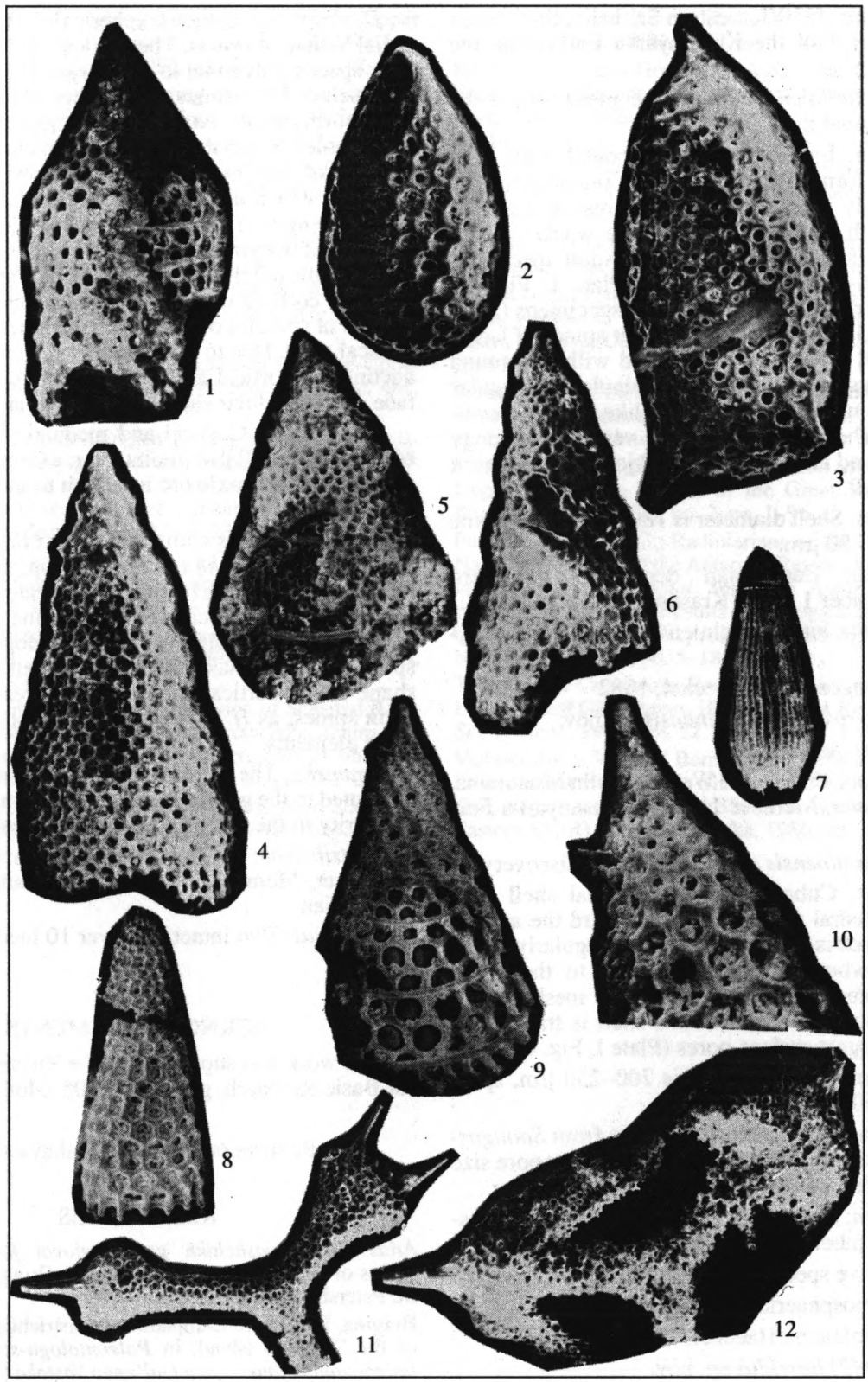


Plate III. (1) *Clathrocyclas* sp., ×200. (2) *Amphipyndax* aff. *ellipticus* Nakaseko et Nishimura, ×200. (3) *Stylodruppa* aff. *bifascicula* Kasinzova, ×500. (4, 5) *Eucyrtidium* sp. A, ×400. (6) *Vistylaria magna* Gorka, ×400. (7) *Archaeodictyomitra squinaboli* Pessagno, ×100. (8) *Cornutella*? sp., ×200. (9) *Clathrocyclas* sp. A, ×200. (10) *Clathrocyclas* sp. A, ×500. (11) *Patulibracchium vereshagini* Kasinzova, ×200. (12) *Patulibracchium* cf. *petroleumensis* Pessagno, ×500.

Holotype: no. 4850/1, western Sakhalin, the Susuya River, Member 1 of the Krasnoyarka Formation, the Campanian.

Name: *sachalinica* points to the place where the species was found first.

Description. Large biconvex discoidal shell of a subtriangular form. Two pylomes are oriented toward the most closely spaced apices and represent two elongate tubes with peripheral spines. The weakly developed keel has spinules at the apices. Adult specimens have the well-developed pylomes (Plate I, Fig. 1), which are poorly developed in young specimens (Plate I, Figs. 2 and 3), and few external joint spines of larger size. The shell surface is ornamented with fine round pores equal in size and with small spinules of irregular form. Several united spinules look like weakly developed ridges. The shell is filled with very loose spongy tissue condensed into a subdiscoidal joint in the central part of the shell.

Dimensions. Shell diameter is 180–230 μm; pylome diameter is 40–80 μm.

Distribution: Campanian deposits of western Sakhalin, Member 1 of the Krasnoyarka Formation.

Material: Six intact specimens and numerous fragments.

Genus *Spongostaurus* Haeckel, 1882

Spongostaurus (?) sachalinensis sp. nov.

Plate I, Figs. 4a and 4b

Holotype: no. 4850/2, the West Sakhalin Mountains, the Susuya River, Member 1 of the Krasnoyarka Formation.

Name: *sachalinensis* after the place of discovery.

Description. Cuboid, almost spherical shell with four short trihedral spines oriented toward the angles. The shell surface is pierced with oval to regularly round large pores, whose diameter is equal to the space between neighboring pores. The spongy meshwork of the inner cavity is so fine that the shell is translucent through the largest surface pores (Plate I, Fig. 4a).

Dimensions. Shell diameter is 200–250 μm, spine length is 40–60 μm.

Comparison. The species is distinct from *Spongostaurus (?) hokkaidoensis* Taketani by a larger pore size and the character of meshwork.

Distribution: the West Sakhalin Mountains, the Susuya River, Member 1 of the Krasnoyarka Formation.

Material: five specimens.

Family Cubosphaeridae Haeckel, 1882

Genus *Hexalonche* Haeckel, 1882

Hexalonch (?) horridus sp. nov.

Plate II, Figs. 1, 2, and 4–6.

Holotype: no. 4850/3, the West Sakhalin Mountains, the Susuya River, Member 1 of the Krasnoyarka Formation.

Name: *horridus* (Latin.) means horrible.

Description. Spinous spherical form with robust radial Y-shaped spines. The cortical shell shows irregularly spaced polygonal to oval large pores, the diameter of which is 2 to 3 times greater than the space between neighboring pores. At the pore conjunction sites, there are spinules of variable length and thickness. The inner surface of the cortical shell is covered with thin spinules, which are similarly disposed but more variable in length. This surface is not ideally spherical because of varying thickness. The medular spherical shell is almost 2 times less in diameter being connected with the cortical one by 6–8 main spines and by the longest of spinules ornamenting the inner surface of the cortical shell. Due to the furcate shape of spinules connecting the cortical and medular shells, the outer surface of the medular shell has an irregular form.

Dimensions. Cortical and medular shells are 400–600 and 150–200 μm in diameter; external parts of the main spines approximate in length to the cortical shell diameter.

Comparison. As compared to *Hexalonche (?) senta* Koslova, *Hexalonche (?) horridus* sp. nov. shows the following distinctive features: (a) greater variability of pore size in the cortical shell; (b) thinner medular shell with different porosity; (c) more developed and furcate spinules, which make the medular shell as irregular in shape as the cortical shell; and (d) greater number of main spines, as *H. (?) senta* has not more than five of these elements.

Remarks. The species described is conventionally attributed to the genus *Hexalonche* because of the close similarity in the outer and inner structure.

Distribution: the West Sakhalin Mountains, the Susuya River, Member 1 of the Krasnoyarka Formation; Campanian.

Material: Two intact and over 10 incomplete specimens.

ACKNOWLEDGMENTS

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Correlation of Upper Mesozoic Sedimentary and Volcanogenic Deposits of Korea and Their Depositional Settings

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Abstract—The Upper Mesozoic deposits of the northern and southern parts of the Korean Peninsula significantly differ from each other. The Upper Jurassic–Neocomian deposits of North Korea are represented by the differentiated volcanics and tuffaceous–terrigenous rocks. In the south, they are composed of the discretely distributed continental terrigenous sediments. The upper Hauterivian–middle Albian deposits bear features in common throughout the Korean Peninsula. They are of the continental terrigenous type (with several cycles of sedimentation), and their thickness sharply increases in the southern part of the Korean Peninsula, where, in addition, basaltoids of the intraplate type occur. In North Korea, the upper Albian–Upper Cretaceous deposits are represented by continental terrigenous rocks of a limited distribution range, whereas in South Korea, by a thick volcanogenic–plutonic basic-to-acid rock series. Settings of the Late Mesozoic sedimentation in the Korea–Japan region are reconstructed. In the Late Jurassic–Neocomian, the lateral structural succession included: (1) the North Korean marginal volcanogenic belt, (2) a system of fore-arc troughs (South Korea and Inner zone of Japan), and (3) deep-sea trench (the boundary between Inner and Outer zones of Japan). Subduction setting also existed at the late Albian–Late Cretaceous stage, however the lateral succession of structures (including the Korean–Japanese volcanic belt) was displaced toward the paleo-Pacific. The late Hauterivian–middle Albian period was marked by cessation of subduction, by intense crustal compression in the continental margin, and by formation of intermontane depressions bounded by faults.

Key words: *Korea, Japan, Upper Mesozoic, correlation, deposition settings, lateral successions.*

INTRODUCTION

Inasmuch as the Korean Peninsula represents, in its geographical position, a natural transsect across the eastern margin of the Asian continent, the study of the Upper Mesozoic deposits within this region allows, first, regularities of changes in their composition occurring toward the Pacific Ocean to be established and, second, the Late Mesozoic geological evolution of the considered section of the Pacific continental borderland to be traced. It should be noted, however, that in the last decades there were different approaches to the study of geological structure of the northern and southern parts of the Korean Peninsula, that created significant difficulties in the correlation of Upper Mesozoic deposits of this region. Our studies carried out in the northern (1986–1990) and southern (summer of 1997) parts of the Korean Peninsula helped to overcome, to a large extent, these difficulties. It should be emphasized that these investigations were conducted together with geologists from the Geological Institute of Academy of Sciences of the Korean People's Democratic Republic and from the Kyungpook National University of North Korea. We also took into consideration various geological data obtained by researchers from these scientific bodies during last decades.

Provided that the Asian continental margin comprised in the Late Mesozoic not only the territory of the Korean Peninsula, but also that of Japan, the summarizing part of the article includes the comparative analysis of the Upper Mesozoic deposits of the entire Korean–Japanese region.

UPPER MESOZOIC DEPOSITS OF NORTH KOREA

The Upper Mesozoic rocks in the northern part of the Korean Peninsula are of a limited distribution range (Fig. 1) and referred to three stratigraphic intervals: Upper Jurassic–Neocomian, upper Hauterivian–Albian, and undifferentiated Lower–Upper Cretaceous (Fig. 2).

The Upper Jurassic–Neocomian. Undifferentiated Upper Jurassic–Neocomian deposits are sporadically exposed in many areas of North Korea. Almost universally, they are bounded by normal or strike-slip faults of the postsedimentary origin in most cases (Filatova, 1990). The faults should be taken into consideration during stratigraphic investigations, because the block structure often creates a false impression of multiple alternation of lithologically similar members in the sec-

tion and of their extremely large total thickness (*Geologiya Korei*, 1993).

The Upper Jurassic–Neocomian deposits of North Korea are united into the Jason Group of a complex terrigenous–volcanogenic composition, though it preserves a regular rock succession in every distribution area, and this property makes correlation easier.[1] In different publications, however, coeval units are traditionally termed differently (*Geologiya Korei*, 1993; and others).

One of the most complete sections of the Jason Group exposed in the Taedongang River basin, near Pyinguang (Fig. 1, Site 4) includes three formation: Chimchon, Taebosan, and Mangendae (Figs. 2, 3). The lower, Chimchon Formation (300–400 m thick) overlies unconformably the Lower Jurassic Sonnimsan Formation and is composed of terrigenous and pyroclastic rocks of pink–gray, red–brown, and violet coloration. Most abundant rocks are represented by sandstones, siltstones, conglomerates composed of angular–rounded clasts of local rocks (Precambrian and Paleozoic gneiss and schist, less commonly sandstone, siltstone, and coal of the Sonnimsan Formation). Mylonitized rocks from the Sonnimsan Formation also occur among clasts, suggesting tectonic movements that occurred directly prior the deposition of the Chimchon Formation.

The most characteristic lithological feature of the Chimchon Formation is the presence of psephitic and psammitic tuffaceous sandstones and tuffites enclosing the uneven pyroclastic fragments of basalts, andesites, and rare dacites.

The younger, Taebosan Formation (800 m) rests, without signs of unconformity, upon the Chimchon Formation and is mainly composed of volcanogenic basic-to-acid rock series. Its basal part includes flows of dark-gray glassy andesites to 170 m thick. Upsection, they are replaced by green and violet medium- and coarse-grained tuffs about 400–600 m thick and displaying mostly andesitic to andesite-dacitic composition. The Taebosan Formation is crowned by rhyolite lava flows of a limited extension and by numerous extrusive domes composed of irregularly crystallized acid rocks ranging in composition from rhyiolites to granite–porphyries and microgranites.

The upper, Mangendae Formation (350 m) overlies without visible unconformity the Taebosan Formation and is represented by alternating light-gray and pink tuffaceous sandstones and gritstones, acidic tuffites and tuffs. Tuffaceous–terrigenous rock varieties prevail; an admixture of the pyroclastic material amounts to 50% in these rocks. The total thickness of the Jason Group in the Taedongang River basin is about 1600 m.

[1] Hereinafter (including figures), we use those units of the regional stratigraphic scale, which are accepted by North Korean, South Korean, and Japanese geologists for corresponding geographic areas.

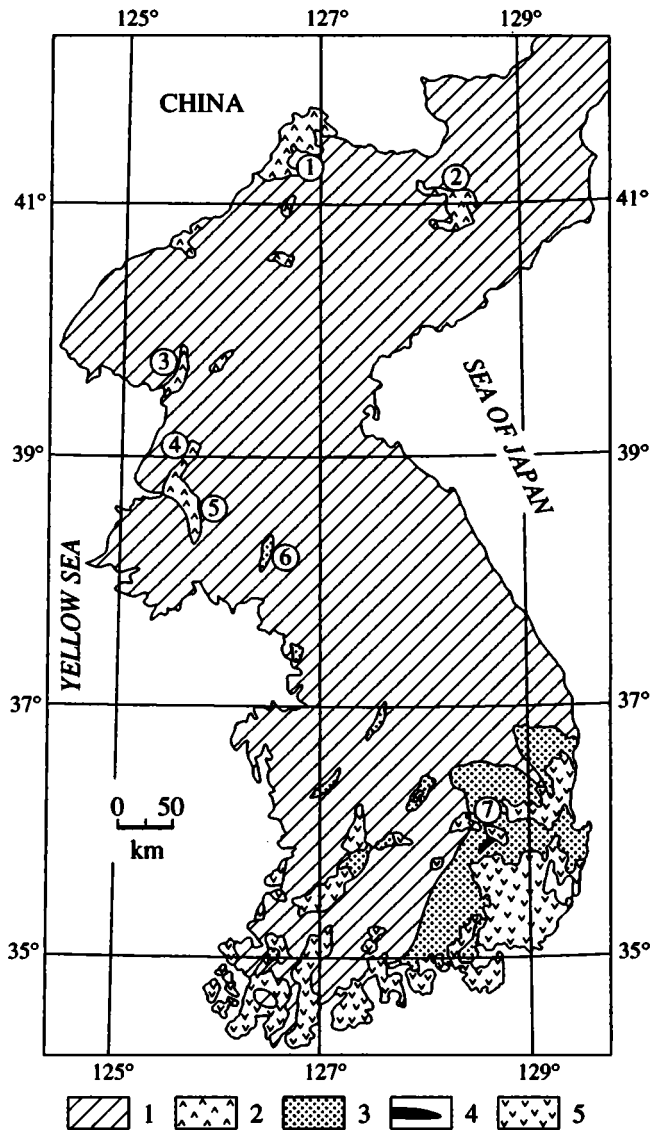

Fig. 1. Distribution of the Upper Jurassic–Paleogene deposits in the Korean Peninsula: (1) pre-Jurassic rocks; (2) Upper Jurassic–Neocomian terrigenous–volcanogenic deposits; (3) upper Hauterivian–Albian terrigenous deposits; (4) Lower Cretaceous subalkaline basaltoids (out of scale); (5) undifferentiated upper Albian–lower Paleogene volcanogenic and intrusive rocks. Numbers in circles correspond to geographic areas of Upper Jurassic–Paleogene rocks: (1) Amnokan River basin, (2) Hochongang, (3) vicinity of Anju, Taerong and Chonchongang rivers basin, (4) Taedongang, (5) Jaerongang, (6) Ryesong, (7) Kyongsang.

To the south, in the lower course of the Jaerongang River (Fig. 1, Site 5), the Chimchon Formation is missing, and the Taebosan Formation overlies unconformably the Lower Jurassic sandstone–conglomerate deposits of the Sonnimsan Formation. The Taebosan Formation is represented there by differentiated volcanics ranging from the moderately basic rocks in its lower part to moderately acid in the upper. Rhyiolites

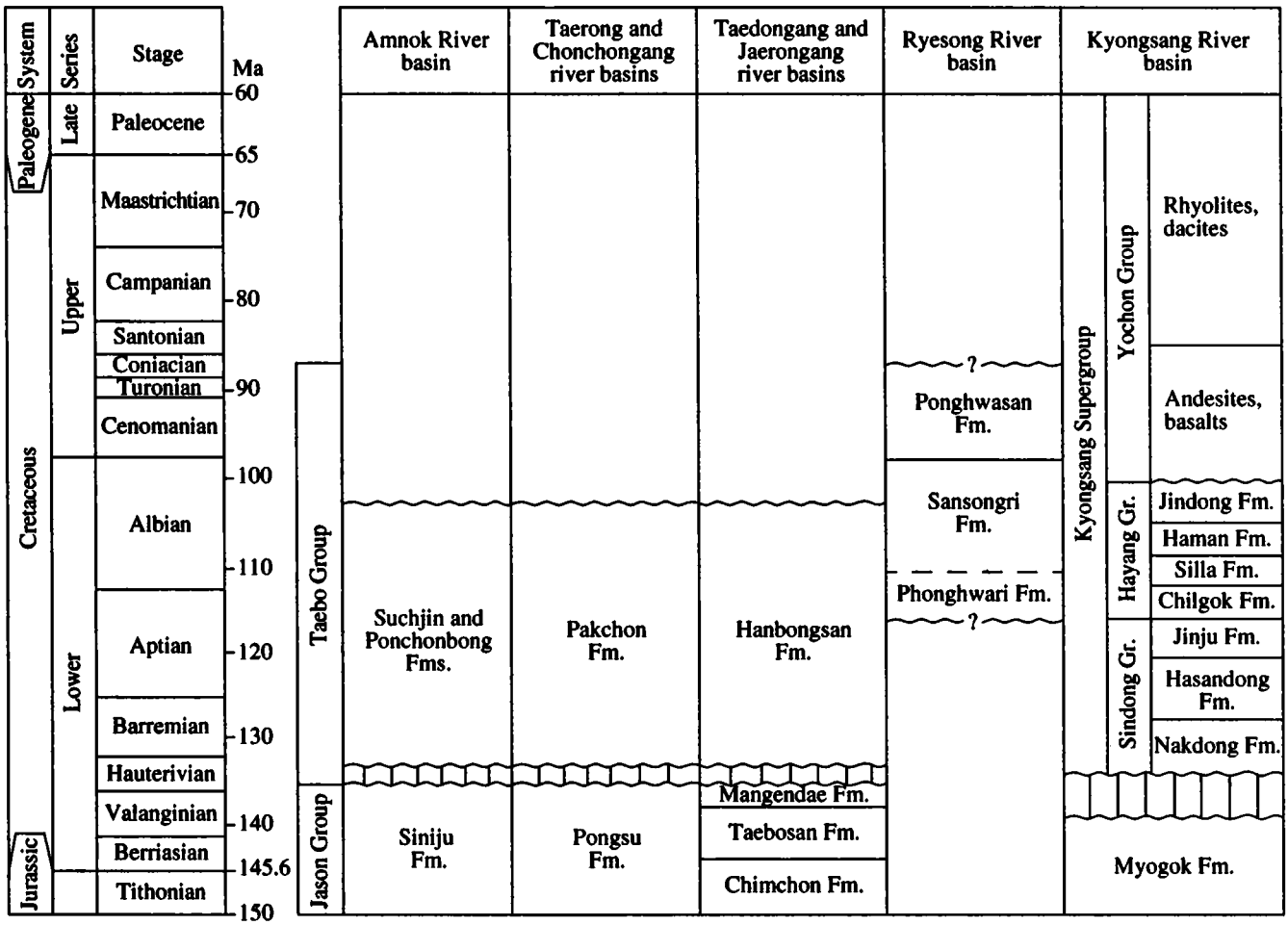

Fig. 2. Correlation of Upper Mesozoic deposits in the northern and southern parts of the Korean Peninsula. Abbreviations: (Fm) formation, (Gr) group.

and their tuffs, as well as tuffaceous–terrigenous varieties, are rare. The overlying Mangendae Formation is limitedly distributed in this area.

Thus, the Jason Group in basins of the Taedongang and Jaerongang rivers is rather uniform, mostly representing the three-member sequence with the main part of volcanics concentrated in its middle part. However, in the north, in the lower course of the Taerong River (Fig. 1, Site 3), the Upper Jurassic–Neocomian deposits were subjected to significant block movements, which feature makes their subdivision difficult. In this area, the single Pongsu Formation 600–900 m thick is established (*Geologiya Korei*, 1969, 1993) as a unit of the three-member structure (Fig. 2). Its lower part consists of gray and yellowish gray conglomerates, sandstones, and siltstones with an admixture of pyroclastic material. The middle part includes dark-green and gray–violet basalts and basaltic andesites with subordinate amount of volcanics (Filatova *et al.*, 1990). The upper part of the Pongsu Formation contains tuffaceous–sedimentary rocks including acidic tuffs.

Northward, in the Amnok River basin (Fig. 1, Site 1), the Jason Group is represented by a thick (not less than 2000 m) tuffaceous–terrigenous and volcanogenic rock succession united into the Siniju Formation (*Geologiya Korei*, 1993) also characterized by the three-member structure (Fig. 2). Its lower part is formed by variegated (red, gray, green) conglomerates, sandstones, and siltstones with rare thin coal seams and pyroclastic andesite fragments. The middle part of the section includes basaltic andesites and andesites in the lower portion of the member and lava–extrusive association of rhyolites and granite–porphyry in its upper portion. The upper part of the Siniju Formation is composed of alternating gray and green tuffaceous sandstones, siltstones, and conglomerates with rare coal seams.

The similar succession of rocks in the Jason Formation is recorded in the Hochongang River basin, near Kapsan (Fig. 1, Site 2).

The Late Jurassic–Neocomian age of the Jason Formation was determined based on findings of freshwater mollusks and plant remains, which are particularly

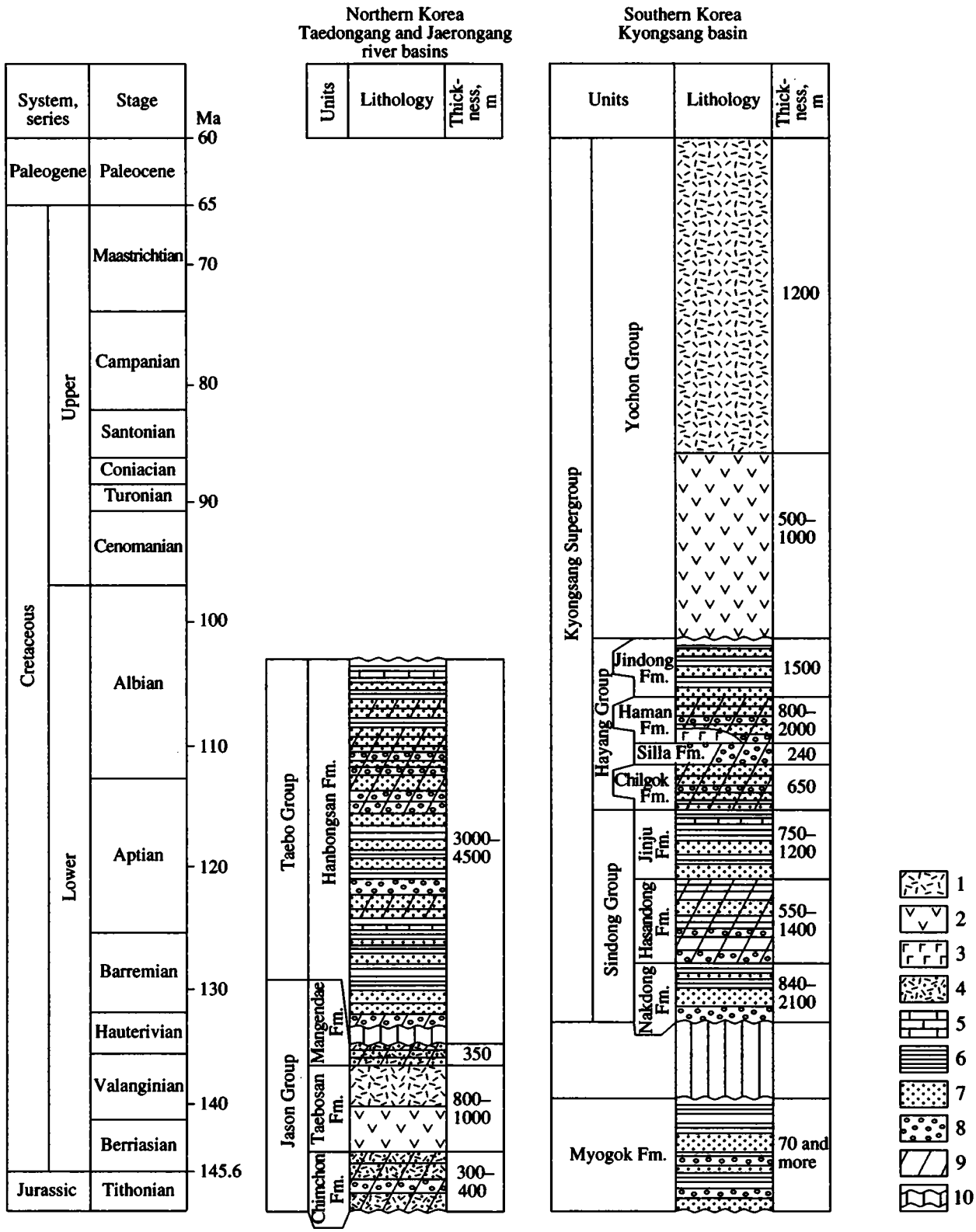

Fig. 3. Correlation of Upper Jurassic–Cretaceous deposits in the northern and southern parts of the Korean Peninsula: (1) rhyolites and tuffs; (2) andesites and tuffs; (3) alkaline and subalkaline basalts; (4) intermediate and acidic tuffs; (5) limestones; (6) agillites, siltstones; (7) sandstones; (8) conglomerates, sandstones; (9) red rocks; (10) unconformity and hiatus; abbreviation (Fm) is for formation.

abundant in the Siniju Formation in the Amnok River basin. The rocks contain *Estheria*, characteristic of the Upper Jurassic and Lower Cretaceous deposits of Mongolia, the Transbaikal region, eastern and northeastern China, and the Primor'e region (Jon, 1987a, b). Abundant plant remains sampled from the Jason Group in basins of the Taedongang, Jaerongang, and Amnok rivers also indicate the Late Jurassic–Early Cretaceous age of host rocks (*Geologiya Korei*, 1993; Pak, 1975, 1984). In the opinion of E.L. Lebedev, a paleontologist from the Geological Institute, Russian Academy of Sciences, the upper age limit for this plant assemblage is the early Neocomian (oral communication).

The upper Hauterivian–Albian. Deposits of this age occurring in North Korea are referred to the Taebo Group and include only sedimentary rocks. They overlie unconformably the Upper Jurassic–Neocomian Jason Group (Fig. 2) and are distributed very discretely (Fig. 1).

In the basins of the Taedongang and Jaerongang rivers (Figs. 2, 3), the Lower Cretaceous deposits of the Taebo Group are united into the Hangbongsan Formation 3000–4000 m thick. Resting above the erosional surface of all older strata, it includes abundant fragments of reworked volcanics from the Jason Group, which sometimes are erroneously considered as an admixture of the pyroclastic material (*Geologiya Korei*, 1993). The lower part of the Hangbongsan Formation is composed of red and gray conglomerates, sandstones, and siltstones with limestone interbeds. The middle part begins with conglomerates, which are replaced higher by alternating siltstones and sandstones. The upper part of the formation again includes coarse-grained red rocks.

To the north, in the Taerong River basin (Figs. 1, 2), the Lower Cretaceous Pakchon Formation (2000–2500 m) overlies unconformably the Pongsu Formation and is composed of basal red conglomerates higher replaced by alternating sandstones, siltstones, and conglomerates. The upper part of the Pakchon Formation is represented by alternating gray and dark argillites, siltstones, sandstones, and rare limestones. The thick (about 2000 m) section of the Lower Cretaceous deposits in the Amnok River basin, where they are known under names of the Sujin Formation or Pongch Formation, is characterized there, similarly to other areas of the northern Korean Peninsula, by polycyclic structure: thick conglomerate members occur in basal, middle and upper parts of the section.

The Lower Cretaceous Taebo Formation encloses abundant fish skeletons and shells of gastropod, ostracods, and estherias. Based on the ostracod assemblage, it is dated back to the Hauterivian–Barremian (*Geologiya Korei*, 1993). Abundant plant remains sampled through the entire section allow the enclosed deposits to be referred to the second half of the Lower Cretaceous (*Geologiya Korei*, 1993; E.L. Lebedev, personal communication). Thus, the Hanbongsan and Pongch formations and their stratigraphic analogues correspond in age to the second half of the Early Cretaceous, most probably to the late Hauterivian (or early Barremian)–early Albian.

The undifferentiated Lower–Upper Cretaceous deposits. The Cretaceous deposits of the Ryesong River basin (Fig. 1, Site 6) occupy a peculiar position. They fill the longitudinal graben of the same name and bear imprints of Late Cretaceous plants in their upper part. These deposits are separated from the above-described Lower Cretaceous rocks of the Taebo Group and are spatially close to the Cretaceous rocks exposed in the southern part of the Korean Peninsula. Their section is characterized by a three-member structure (Fig. 2). Two lower formations (Phongwari and Sansongri) are lacking fossils and consist of alternating conglomerates, sandstones, and siltstones. Conglomerates are preponderant in the Phongwari Formation (about 250 m). The overlying Sansongri Formation (about 900 m) is characterized by less coarse rock varieties.

The overlying Upper Cretaceous Ponghwasan Formation (420 m) is composed of alternating gray and violet siltstones and argillites with subordinate sandstone interbeds. The Late Cretaceous age of the formation is established based on the enclosed plant remains (*Geologiya Korei*, 1993; Pak, 1989). In addition, it bears ostracod shells and dinosaur footprints.

The age of two underlying formations occurring together in a continuous section of the Ryesong River basin is unclear, and only indirect evidence can be considered. It is difficult to agree with the reference of the Ponghwari and Sansongri formations to the Jason and Taebo groups, respectively, as is accepted in some publications (*Geologiya Korei*, 1969, 1993), because the groups are separated by an unconformity in the Korean Peninsula. In addition, both formations are barren of volcanics characteristic of the Jason Group. The absence of volcanic rocks and abundance of red deposits allow these formations to be correlated with the Taebo Group of North Korea and with the Hayang Group of South Korea (Fig. 2). Most likely, the Ponghwari and Sansongri formations correspond to a part of the Hanbongsan Formation (and its stratigraphic analogues), and span the time range of the second half of the Early Cretaceous. It is not inconceivable, however, that all three formations occurring in the Ryesong graben are of the Late Cretaceous age.

UPPER MESOZOIC DEPOSITS OF SOUTH KOREA

In the southern part of the Korean Peninsula, Upper Mesozoic deposits occupy larger areas than in its northern part (Fig. 1). In addition to small, structurally isolated blocks, they cover a spacious southeastern part of the peninsula, which is usually termed as the Cretaceous Kyongsang basin. Owing to the block structure, this basin is subdivided into three subbasins with

slightly different sedimentary sections (Chang, 1987). With respect to stratigraphy, the Upper Mesozoic deposits of the southern part of the Korean Peninsula are referred to three intervals: Upper Jurassic–Lower Cretaceous, upper Hauterivian–middle Albian, and upper Albian–Upper Cretaceous.

The Upper Jurassic–Lower Cretaceous. This interval includes the Myogok Formation (Figs. 2, 3) exposed in small erosional "windows" along the northern border of the Kyongsang basin. Rocks of this formation are deformed more intensely as compared with the unconformably overlying Lower Cretaceous strata of the basin. In its lower part, the Myogok Formation is composed of alternating siltstones, coarse-grained sandstones and conglomerates with coal seams. Higher in the section, these rocks are replaced by black and dark-gray argillites, siltstones, and fine-grained sandstones. This formation is often referred to the Upper Jurassic based on finds of the respective freshwater mollusks (Chang, 1987; Um, 1979). However, abundant plant remains from these rocks are characteristic of both the Late Jurassic and Early Cretaceous assemblages. Thus, the Late Jurassic–Early Cretaceous (Neocomian) age of these deposits is not inconceivable.

The upper Hauterivian–middle Albian. The Lower Cretaceous deposits of the Kyongsang basin are divided into the Sindong and Hayang groups (Figs. 2, 3) (Chang, 1987; Chang *et al.*, 1990, 1997; Chang and Park, 1995; Chang and Son, 1994). The Sindong Group includes three formations. The lower Nakdong Formation (Fig. 3) overlies unconformably the deformed layers of the Myogok Formation and contains basal boulder–pebble conglomerates, which are replaced higher by alternating sandstones, siltstones, and argillites including the carbonaceous varieties of the latter.[2] Such a structure of the Nakdong Formation allows it to be considered as representing an autonomous stage of sedimentation (Chang and Son, 1994). The overlying Hasandong Formation is characterized by abundant red rocks. Coarsest varieties (including conglomerates) are confined to its lower part, whereas sandstones and siltstones compose the upper part of the section. The Sindong Group is crowned by the Jinju Formation that is composed of gray-colored, relatively fine-grained rocks with prevailing black siltstones, argillites, and single limestone interbeds in its upper part (Fig. 3). The integral section of the Hasandong and Jinji formations reflects an autonomous sedimentation cycle (Fig. 3) (Chang and Son, 1994).

The younger Hayang Group includes four formations (Fig. 2). The lower red Chilgok Formation is considered as representing a polycyclic succession that was formed during several sedimentation stages (Chang and Son, 1994). Its lower part (230 m) is composed of alternating siltstones, argillites, and sandstones. The middle part (210 m) is formed by alternating sandstones and siltstones with conglomerate interbeds. Pebble in conglomerates is represented by andesites, dacites, rhyolites, tuffs of the same composition, quartzites, and granites. In the upper part of the formation (more than 200 m), abundant conglomerate interbeds occur again alternating with sandstones and siltstones. The overlying Silla Formation is used as one of the most distinct reference units in the Kyongsang basin because it is composed of thick (240 m) red conglomerate beds. In addition to quartzite, gneiss, sandstone, and hornfels, the Silla conglomerates contain abundant fragments of volcanic rocks variable in composition, as well as granite–porphyries and diorite–porphyries. The younger, Haman Formation is mainly composed of alternating red sandstones, siltstones, and marls with subordinate conglomerate interbeds. The peculiar feature of this formation is the packet of the Hakbong subalkaline and alkaline basalt flows (Fig. 3). They spread approximately over 15 km and are to 200 m thick. The Hakbong Basalts show, similarly to those formed in the intraplate setting, the elevated and moderate K_2O and TiO_2 contents and are nepheline-normative. According to Nb and Y contents, they are intermediate between island-arc and intraplate basalts (Kim, 1993).

The Jindong Formation, the youngest one in the Hayang Group, is formed by dark-gray and black shales, siltstones, and sandstones terminating the sedimentation cycle that began with conglomerates of the Silla Formation.

The total thickness of the Sindong and Hayang groups varies from 7000 to 10000 m within the Kyongsang basin. Both of them contain abundant freshwater mollusks and plant remains (Chang, 1987; Um, 1979; and others). In addition, there are dinosaur bones, eggs, and footprints occurring throughout their entire sections (Chang, 1987). However, all these fossils are useful for interpreting environments rather than for age determination. Nonetheless, finds of charophytes and pollen–spores assemblages, as well as the isotopic age of volcanics from the overlying Yochon Group allow the Sindong and Hayang groups to be dated back to the late Hauterivian–middle Albian. Accumulation of the Sindong Group commenced, most likely, at the end of the Hauterivian (Chang, 1994), because the upper part of the Nakdong Formation bears charophytes (Seo, 1985) characteristic of the Hauterivian and early Barremian (Wang and Lu, 1982). Rocks from the upper part of the Hayang Group yield a palynological assemblage characteristic of the Aptian and early Albian (Choi, 1985). Inasmuch as the oldest volcanics of the younger Yochong Group are dated back to the late Albian (*Isotope Age ...*, 1995a, b), the upper part of the Hayang Group most likely should be referred to the middle Albian.

The upper Albian–Upper Cretaceous. In South Korea, this interval is represented by a thick sequence

[2] In the South Korean literature, dark-gray and black alluvial and lacustrine siltstone and argillites of the Kyongsang basin are usually termed as "shale" or "black shale."

of volcanogenic rocks (belonging to the calc-alkaline series) united into the Yochon Group, which overlies unconformably all the oldest deposits including the Aptian–middle Albian strata (Figs. 2, 3). They form, together with the Pulguksa granites, a common volcano–plutonic rock association (Jin, 1988). Although the volcanics of the Yochon Group are related to the development of different strato-volcanoes (Hwang, 1996; Hwang and Kim, 1994; Kim *et al.*, 1991, 1993; Won *et al.*, 1994), they are universally characterized by the two-member structure. The lower part of the section is composed of basalts, andesites, and their tuffs (often coarse-grained), whereas the upper layers are represented by dacites, rhyolites, and acidic tuffs occurring in association with abundant extrusive domes. In chemical properties (Hwang and Kim, 1994; Jin, 1988; Won *et al.*, 1994), these rocks are close to volcanics of marginal belts. Their enrichment in potassium and incoherent elements is mist likely explained by the confinement of volcanic sequences of South Korea to the rear part of the volcanic belt whose frontal fragments are located in Japan. The high initial $^{87}Sr/^{86}Sr$ ratio (from 0.7055 to 0.7244) (Hwang, 1996; Jin, 1988) indicates probably that volcanics are related in origin to both the mantle and crustal sources.

According to isotopic data (*Isotope Age...*, 1995a, b), rocks of the Yochon Group are of the terminal Early Cretaceous–Paleocene age. The results of the Rb-Sr and K-Ar analyses of some minerals (apatite, zircon, biotite, potassic feldspar) and whole-rock samples reveals two peaks of volcanism: at the Albian–Cenomanian boundary (97–90 Ma) and in the Campanian–Maastrichtian (83–65 Ma). The first of these pulses is marked by eruptions of intermediate–basic volcanics and the second one by acidic volcanic extrusions. The latter ceased in the period of 65–60 Ma (early Paleocene), although the bulk of volcanics of the Yochon Group was formed by the end of the Maastrichtian.

CORRELATION OF UPPER MESOZOIC DEPOSITS IN NORTHERN AND SOUTHERN PARTS OF THE KOREAN PENINSULA

Comparison of the Upper Jurassic–Cretaceous deposits in northern and southern parts of the Korean Peninsula reveals their essential differences, which are particularly distinct for the Upper Jurassic–Neocomian and upper Albian–Upper Cretaceous intervals (Fig. 3).

In the northern part of the peninsula, the upper Jurassic–Neocomian sequence is composed of terrigenous–volcanogenic deposits, which were previously defined as representing rocks of the North Korean marginal belt (Filatova 1990; Filatova *et al.*, 1990) The three-member section of the Jason Group (Fig. 3) reflects the three-stage cycle of the development of this volcanic belt: from terrigenous and pyroclastic rocks of the initial stage (Chimchon Formation) to thick volcanogenic deposits of the middle stage (Taebosan Formation) and to sedimentary–terrigenous deposits of the terminal stage (Mangendae Formation). In petrologic and geochemical properties (including REE distribution patterns), the differentiated rock association of the Taebo Formation (basalts–andesites–dacites–rhyolites) belonging to the calc-alkaline series is typical of marginal suprasubduction belts (Filatova, 1995). The current fragmentary distribution of the Upper Jurassic–Neocomian deposits in the northern part of the Korean Peninsula originated in the course of subsequent tectonic deformation and movements along strike-slip faults (Filatova, 1990).

In the southern part of the Korean Peninsula, the Upper Jurassic–Neocomian deposits barren of volcanics are represented by the thin terrigenous Myogok Formation (Fig. 3). The deposits of the next, late Hauterivian–middle Albian sedimentation stage show many features in common throughout the entire peninsula. They overlie Upper Jurassic–Hauterivian deposits with a hiatus probably spanning the major part of the Hauterivian. These deposits that were formed in fluvial and, less commonly, in lacustrine settings are terrigenous in composition; they universally contain pebbles and boulders of the Upper Jurassic–Neocomian volcanics (*Geologiya Korei*, 1993; Filatova *et al.*, 1990; Chang *et al.*, 1997).

The Upper Hauterivian–Albian section is characterized by the polycyclic structure distinguishable in the entire peninsula. In the northern part, the section reveals three large sedimentation cycles. The first cycle begins with conglomerates of the Hanbongsan Formation and terminates with the Barremian dark-colored argillites, siltstones, and sandstones intercalated with limestone interbeds (Fig. 3).

The next, Aptian sedimentation cycle also begins with red conglomerates and terminates with a member of alternating gray and greenish gray sandstones and siltstones. The thickest conglomerate member corresponds to the beginning of the third, Albian cycle, which is terminated with black and dark-gray siltstones, argillites, sandstones, and limestones accumulated in the lacustrine setting. These three sedimentation cycles correspond to three episodes of orogenic movements, which were most intense in the Hauterivian, terminal Aptian, and initial Albian, as is evident from the relative conglomerate thickness. The similar polycyclic structure is also characteristic of the upper Hauterivian–Albian terrigenous deposits in the southern part of the Korean Peninsula, although they are locally twice thicker (to 10000 m) (Chang *et al.*, 1990, 1994). Conglomerates enclosed in these deposits are also characterized by a greater thickness. The section in question consists of three large cycles (Fig. 3). The first, shortest cycle begins with basal conglomerates of the Nakdong Formation and terminates with fine-grained rocks belonging to the same unit. Thick red conglomerates and coarse-grained sandstones of the Hasandong Formation form the base of the second sedimentation cycle, the end of which was marked by

accumulation of black and gray argillites, siltstones, and limestones of the Jinji Formation. The upper Aptian–Albian Hayang Group corresponds to the third, most durable cycle, which starts with thick red sandstones and conglomerates of the Chilgok, Silla, and Haman formations. It should be noted that pebbles and boulders from the Hayang Group are represented not only by quartzite, gneiss, schist, granite, and volcanics (from basalts to rhyolites) characteristic of the coeval deposits in the north of the Korean Peninsula, but also by abundant fragments of the Permian, Triassic, and Upper Jurassic–Lower Cretaceous jasper and siliceous shales very similar to coeval rocks from Japan (Chang *et al.*, 1995, 1997). The initial stage of the considered cycle was marked by the eruption of alkaline and calc-alkaline lavas (Fig. 3). The terminal stage of the third cycle corresponds to the thick (to 1500 m) dark-gray fine-grained rocks of the Jindong Formation.

The comparison of the Hauterivian–middle Albian deposits from the northern and southern parts of the Korean Peninsula reveals their great similarity. They are non-marine polycyclic terrigenous deposits with red coarse-grained rocks that occur in the section at three levels and higher are replaced by gray-colored fine-grained and chemogenic varieties. In both the northern and southern parts of the Korean Peninsula, the upper Aptian–Albian conglomerates are of the maximum thickness. Their important feature is the presence of reworked Precambrian and Paleozoic metamorphosed and sedimentary rocks derived from the adjacent areas and mixed with the Upper Paleozoic–Mesozoic (including Neocomian) jaspers and siliceous shales similar to those from the Mino–Tanba–Ashio–Chichiby zones of northern and central Japan. Moreover, the upper Aptian–Albian conglomerates from the Kyongsang basins in the southern Korean Peninsula contain abundant clasts of volcanics variable in lithology and transported from the North Korean volcanic belt. All these facts indicate that the most intense Early Cretaceous orogenic movements occurred during the last, late Aptian–Albian stage.

However, there are also some differences between rocks from the northern and southern parts of the Korean Peninsula. They are thicker in the southern part of the peninsula as compared with sections in the northern part. Coarse-grained facies initial in each sedimentation cycle are also significantly thicker in this region. In addition, packets of alkaline and subalkaline basalt flows at the upper Aptian–Albian level occur only in the southern part of the peninsula.

The upper Albian–Upper Cretaceous rocks are essentially different in the northern and southern parts of the Korean Peninsula. Only one small area of the Upper Cretaceous terrigenous deposits is recorded in the north of the peninsula south of the distribution areas of the Lower Cretaceous deposits. In the southern half of the peninsula, the upper Albian–Upper Cretaceous volcano–plutonic association of the Yochon Group unconformably overlying the Lower Cretaceous deposits is widespread as a constituent of the Korean–Japanese marginal volcanic belt.

LATERAL SUCCESSIONS OF LATE MESOZOIC SEDIMENTATION SETTINGS IN THE KOREAN–JAPANESE REGION

The lithologic changes in the Upper Mesozoic deposits of the Korean Peninsula allow the regular lateral succession of structures characteristic of the continent–ocean transition zone to be reconstructed in the region under consideration. Moreover, changes in the structure of the Upper Mesozoic section are indicative of their evolution with time. It is impossible however to understand this evolution without data on the Japanese territory that was, together with the Korean Peninsula, a part of the common eastern margin of the Asian continent in the Late Mesozoic (Taira, 1979, 1985; Taira and Tashiro, 1987; and others).

The Late Jurassic–Valanginian stage was characterized by the formation of the Korean–Japanese marginal volcanic belt (Fig. 4, Column 1) that stretched southward into China (*Atlas ...*, 1985). In the southern part of the Korean Peninsula, only isolated outcrops of terrigenous continental deposits are known. In the northern part of the Inner zone of Japan (the Hida zone, Fig. 4, inset, Site 3'), there is a regressive succession of terrigenous rocks, the lower part of which was formed in the brackish-water lagoonal environments, whereas deposits of the upper part are of the fluvial and lacustrine types (Matsukawa *et al.*, 1997). Closer to the ocean, in the Sangun zone of Inner Japan (Fig. 4, Column 3), the Upper Jurassic–Neocomian terrigenous deposits of the Toyonishi Group were mainly formed in the shallow-water marine environments. The easterly Mino–Tanba–Ashio and Chichiby zones (contiguous regions of the Outer and Inner zones of Japan, Fig. 4, Columns 4 and 5) host deep water turbidites with melange horizons and siliceous rocks, which contain radiolarians of Late Jurassic–Early Cretaceous (including Valanginian) age (Wakita, 1988) and are considered as representing the accretionary complex of the Jurassic–Early Cretaceous subduction zone (Matsukawa *et al.*, 1997),

Thus, the Late Jurassic–Neocomian lateral succession of structures of the Korean–Japanese region included the North Korean marginal volcanic belt, adjacent seaward to a system of fore-arc (fore-belt) troughs formed in the continental margin. Toward the ocean, sedimentation environments in the troughs changed from terrestrial (the southern part of the peninsula) to marine (its northern part). Areas located farther to the east were occupied by a deep-sea trench, where an accretionary complex has been formed. Rocks of the complex are exposed now in the Mino–Tanba–Ashio and Chichiby zones. In the period of 150–135 Ma, these two zones were inside a common sedimentation area, and the Median Tectonic Line separated them

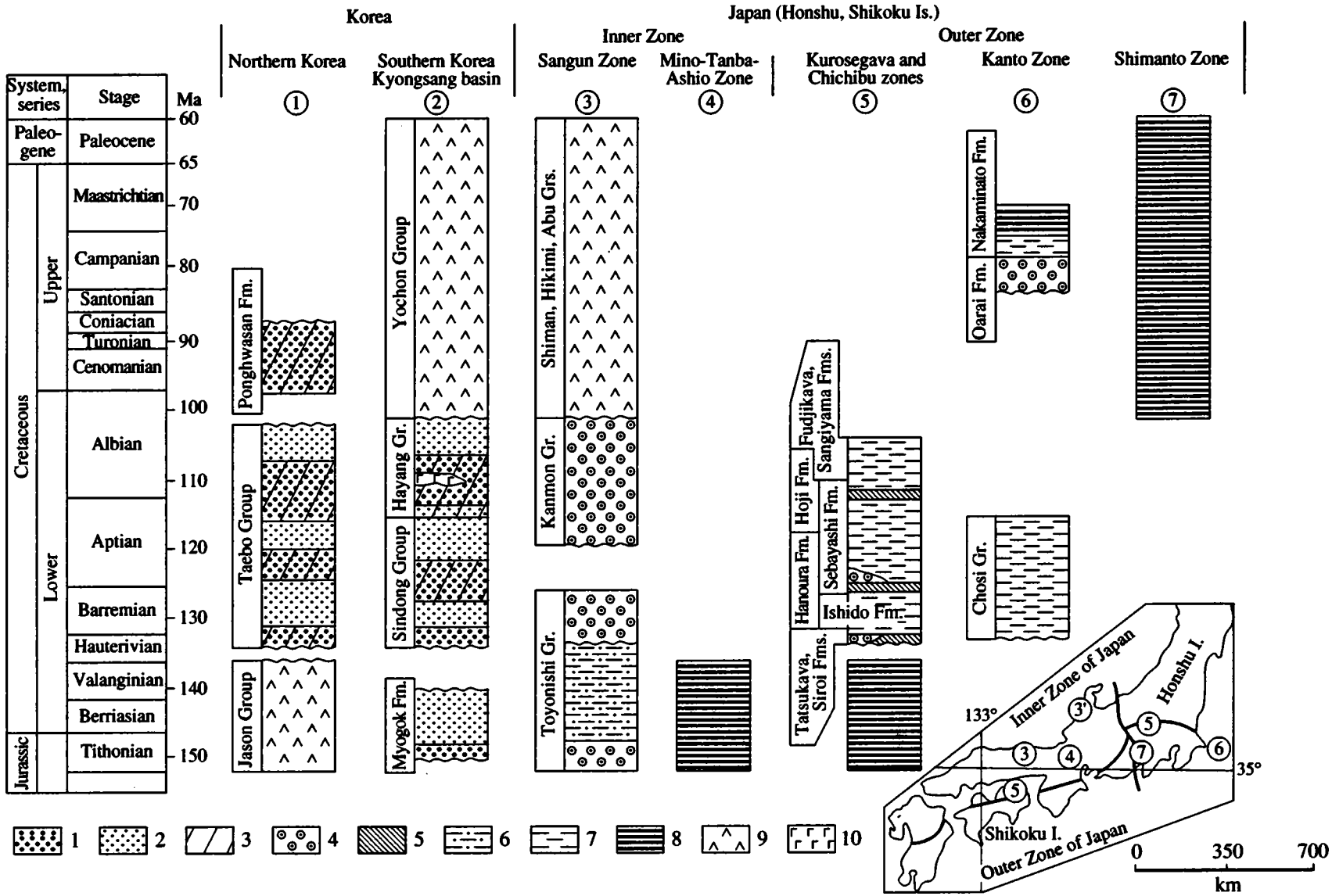

Fig. 4. Correlation of Upper Mesozoic deposits of the northern and southern parts of the Korean Peninsula: (1–3) alluvial and lacustrine deposits: (1) coarse-grained (conglomerates included), (2) fine-grained (black shales and limestones); (3) red rocks; (4) alluvial and lacustrine deposits undifferentiated; (5) brackish-water deposits; (6) coastal and coastal-marine; (7) marine (including shallow-water) deposits; (8) turbidites, siliceous rocks, mélange (complexes of deep-sea trenches); (9) basalt–andesite–dacite–rhyolite association of marginal volcanic belts; (10) alkaline and subalkaline basalts of extention zones; abbreviations: (Fm) formation, (Gr) group.

Inset map shows the location of the considered sections of the Upper Mesozoic deposits in Japan. (Section 3 after Sakai and Okada, 1997; Murakami, 1985, Sections 3', 1, 4, 5, 7 after Matsukawa *et al.*, 1997; Section 6 after Ando, 1997; Matsukawa *et al.*, 1997). The Outer and Inner zones of Japan are divided by the Median Tectonic Line.

later, during the Middle Cretaceous orogenic phase (Matsukawa *et al.*, 1997).

The late Hauterivian–middle Albian stage was marked by the complete change of deposition environments in the Korean–Japanese region. During this period, the North Korean marginal volcanic belt died off, and the belt-related structures, including the subduction zone, ceased to exist. This tectonic reorganization was related to intense orogenic movements, which resulted in significant differentiation and uplift of the Korean–Japanese region and generated the unconformity separating all older strata from the upper Hauterivian–middle Albian deposits.

The late Hauterivian–middle Albian stage was lacking the subduction events and eruptions of corresponding magmatic rocks. Terrigenous facies accumulated in progressively deeper settings toward the ocean (Fig. 4). The stage was marked by sedimentation in isolated intermontane depressions bounded by normal or strike-slip faults (Chang *et al.*, 1994, 1997; Okada, 1997; Sakai and Okada, 1997; and others). Sakai and Okada (1997) believe that the main role in the formation of these basins was that of the sinistral strike-slip faults. The number of basins increased across the Asian margin toward the ocean. In the Korean Peninsula and Outer zone of Japan (the Sangun zone), sedimentation in intermontane depressions occurred under terrestrial conditions (Fig. 4, Column 3) against the background of intensified orogenic movements. The peak of the latter in the terminal Aptian–initial Albian time resulted in the further uplift of the Asian margin and contrast displacements in a system of horsts and subsided intermontane basins. The clastic material was transported into these basins from the entire Korean–Japanese continental margin. In Japan, erosion affected not only the above-mentioned Upper Paleozoic–Mesozoic jaspers and siliceous shales but also serpentinized ultramafic rocks of collision sutures (Asiedu *et al.*, 1997; Arai and Okada, 1991; Hisada *et al.*, 1997). The peak of orogenic and contrast block movements at the Aptian–Albian boundary corresponds with the maximum accumulation of conglomerates and with an episode of intraplate basaltic eruptions.

Closer to the ocean, in the Hida and Sangun zones of Japan (Fig. 4, Column 3), sedimentation occurred in fluvial, lacustrine, and lagoonal environments. For instance, the Kanmon and Toyonishi groups of the Inner zone of Japan include alluvial and brackish-water facies (Sakai and Okada, 1997). To the east, the Asian margin was covered by the shelf sea, and marine sandy-clayey facies accumulated in the Kurosegava and Chichibu zones (Fig. 4, Column 5). It is remarkable that three-stage orogenic pulses detectable in the Korean sections resulted in the accumulation of three horizons different in age and composed of shallow-water and terrestrial deposits in sections of the Shiroi and Sebayashi formations. Coastal and marine sedimentation environments of the shelf sea are also established for the easterly areas of the Inner Zone (the Chosi Group of the Kanto zone, Fig. 4, Column 6).

Thus, the late Hauterivian–middle Albian time interval in the eastern margin of the Asian continent is marked by the lateral facies succession of terrigenous rocks from terrestrial in the Korean Peninsula to terrestrial and lagoonal in the Inner zone of Japan, and to shallow-water marine in the Outer zone. Neither deep-sea trench nor subduction zone were formed during this period, as it is evident from the absence of the accretionary rock complex.

An extremely intense compression of the eastern margin of the Asian continent during the late Hauterivian–Albian stage is manifested along the entire periphery of the paleo-Pacific (Filatova, 1996, 1998; Vaugan, 1995). It ceased volcanism in the North Korean volcanic belt (and terminated subduction of the oceanic plate) and resulted in subsequent uplift and downwarping of the continental margin accompanied by formation of intermontane depressions. The intense compression in the continental marginal zones during the Middle Cretaceous that was accompanied by the continental growth owing to the accretion of numerous terranes of a different nature correlates with the high rate of the earth crust generation in the interval of 120–100 Ma (Larson, 1991). In the Korean–Japanese region, the Middle Cretaceous compression of the continental margin corresponded to its collision with the Kurosegava terrane, which resulted in the formation of a wide zone of sinistral strike-slip faults along the continental margin, including the Median Tectonic Line of Japan (Sakai and Okada, 1997). During this stage, high-magnitude strike-slip faults divided the Late Jurassic–Neocomian North Korean volcanic belt into separate fragments.

The late Albian–Late Cretaceous stage was marked by the principal changes in the lateral succession of structures in the continent–ocean transition zone. During this stage, subduction of the oceanic plate under the eastern margin of the continent regenerated activity in the Korean-Japanese volcanic belt that shifted toward the ocean and occupied the southern part of the Korean Peninsula and northwestern Japan. The back-arc basin, which was located in the rear of the volcanic belt and bounded by faults, accumulated relatively thin terrigenous alluvial and lacustrine deposits. Thick volcanogenic deposits related to the Korean-Japanese volcanic belt occupied the southern areas of the Korean Peninsula and the Inner zone of Japan (Fig. 4, Columns 2, 3). In the latter, the volcanogenic sequence includes upper layers of the upper Albian Simonoseki Subgroup and overlying deposits of the Shiman, Hikimi, and Abu groups (Murakami, 1985; Sakai and Okada, 1997). To the east, the Axial fore-arc trough bounded by faults was located along the Median Tectonic Line of Japan (Sakai *et al.*, 1990; Sakai and Okada, 1997). The trough was filled with sediments of two types: first, with non-marine,

coastal-marine, and shallow-water deposits, and second, with deep-sea terrigenous and flyschoid facies. There existed probably a system of fore-belt troughs, including that of the Kanto zone (Fig. 4, Column 6). These structures host a transgressive succession from the Campanian alluvial conglomerates to the marine (including coastal-marine) sandy–silty sediments and Maastrichtian turbidites. Eastward, the system of fore-arc troughs with continental and marine sedimentation was in contact with the deep-sea trench of the Shimanto zone (Fig. 4, Column 7), where thick upper Albian–Upper Cretaceous turbidites accumulated (Matsukawa *et al.*, 1997). Thus, there was the complete lateral succession of the late Albian–Late Cretaceous structures of the continent–ocean transition zone: the back-arc basin, the marginal volcanic belt (with continuation to the north, in Northeastern Russia and to the south, in China), a system of fore-belt troughs, and the deep-sea trench. Inasmuch as the continental margin in this region grew during the late Hauterivian–middle Albian period, the considered succession of late Albian–Late Cretaceous subduction-related structures shifted toward the ocean relative to the similar succession of structures of the Late Jurassic–Neocomian stage.

CONCLUSION

(1) The Upper Mesozoic deposits of the northern and southern part of the Korean Peninsula reveal significant differences in the Upper Jurassic–Neocomian and upper Albian–Upper Cretaceous sequences being similar in lithology of the upper Hauterivian–Albian sequences. The Upper Jurassic–Neocomian deposits of northern Korea are represented by differentiated volcanics (from basalts to rhyolites) and also by tuffaceous–terrigenous rocks mainly confined to the upper and lower parts of the section. In the southern part of the Korean Peninsula, this stratigraphic interval is composed of locally distributed terrestrial terrigenous deposits. The upper Hauterivian–middle Albian deposits universally rest unconformably upon the older strata and show many features in common throughout the Korean Peninsula. They are of the terrestrial terrigenous type, and their section comprises three large sedimentation cycles beginning with red conglomerates and crowned by black siltstones and argillites with limestone interbeds. The lateral variations of these deposits show a sharp increase of their thickness in the southern areas of the peninsula, where, in addition, basalts of the intraplate type occur. The upper Albian–Upper Cretaceous interval is represented by the local field of continental terrigenous rocks in the northern part of the Korean Peninsula and by a thick volcanogenic–plutonic basic-to-acid rock series in the south.

(2) Correlation of the Upper Mesozoic deposits within the Korean–Japanese region allows the lateral successions of structures in the transition zone between the Asian continent and paleo-Pacific to be reconstructed and sedimentation settings to be interpreted for three age levels.

(a) The Late Jurassic–Neocomian stage was marked by development of the North Korean marginal suprasubduction volcanic belt and a system of fore-arc troughs with the sedimentation environment changing from fluvial and lacustrine (Southern Korea) to lagoonal, coastal-marine, and shallow-water marine (the Inner zone of Japan). This succession of structures in the continental margin was terminated by a deep-sea trench filled with turbidites and mélange of the Mino–Tanba–Ashio and Chichibu zones of Japan.

(b) The Upper Hauterivian–middle Albian terrigenous sequences reflect the lateral succession from terrestrial deposits in the Korean Peninsula to terrestrial and lagoonal sediments in the Inner zone of Japan, and to shallow-water marine facies in the Outer zone. This stage was marked by cessation of subduction, by an intense compression of the continental margin under conditions of collision of the latter with the Kurosegava terrane, and by formation of a system of subparallel basins bounded by faults. This Middle Cretaceous orogeny, paroxysm of which occurred in the terminal Aptian–initial Albian, resulted in tectonic disintegration of the North Korean volcanic belt.

(c) In the late Albian–Late Cretaceous, the Korean–Japanese suprasubduction volcanic belt with a system of fore-arc (Inner and Outer zones of Japan) and back-arc (northern part of the Korean Peninsula) troughs developed over the accreted continental margin; in the east, this system joined the deep-sea trench (turbidites and melange of the Shimanto Zone).

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Pliocene–Quaternary Planktonic Foraminiferal Datums in DSDP Site 397 and Comparison of Zonal Stratigraphic Schemes

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Abstract—Planktonic foraminifers from the Pliocene and Quaternary sediments penetrated at Site 397 of the Deep See Drilling Project (DSDP) were thoroughly studied, and the ages of 20 first and last appearance datums (FAD and LAD) of different species were calculated. They were correlated with the corresponding events from two other climatic belts of the Atlantic. Five reliable biostratigraphic datums were recognized: the first occurrence levels of *Globorotalia margaritae* and *G. crassaformis viola*, and the extinction levels of *Sphaeroidinellopsis seminulina, Globoquadrina altispira*, and *Globorotalia miocenica*. The zonal subdivision of Pliocene–Quaternary deposits corresponds to a somewhat modified zonation of Bolli and Saunders (1985). The latitudinal limitations to apply the scheme in the North Atlantic are outlined. Five taxa that appeared in Site 397 sediments earlier than in more southern and northern Atlantic regions are revealed.

Key words: *Pliocene, Pleistocene, planktonic foraminifers, zonal stratigraphy, first and last appearance datums, stratigraphic zonations, Atlantic.*

INTRODUCTION

In the past three decades, stratigraphic research of deep-sea sediments has culminated in the development of different biostratigraphic zonation schemes for the tropical (Blow, 1969; Bolli and Premoli Silva, 1973) and subtropical (Berggren, 1977) realms, the Mediterranean (Cita, 1973, 1975), and the middle (Berggren, 1972; Weaver and Clement, 1986) and high (Spiegler and Jansen, 1989) latitude areas. These zonations are based either on the same or on different planktonic foraminiferal zonal forms and are sufficiently representative of the microfaunal changes in a certain climatic belt. However, the correlation of high and low latitude sediments using different zonations presents substantial difficulties. Additionally, the occurrence of some tropical zonal forms in subtropical and temperate latitudes and the sections with a transitional composition of microfauna enable the application of the detailed Cenozoic low latitude zonal scheme of Bolli (Bolli and Saunders, 1985) in the subtropical and temperate belts, i.e., the use of a single stratigraphic scheme in most part of the Atlantic. Such a possibility was demonstrated by using the Pliocene–Quaternary deposits of the Rockall Plateau as an example (Krasheninnikov and Bylinskaya, 1994).

Changes in planktonic foraminiferal assemblages within a section is a basis for the zonal subdivision of sediments, and boundaries between major units are determined by the alteration of whole assemblages, but not of certain species. However, in recent years a great deal of papers provide data on the age of first and last occurrence levels of species important for stratigraphy that can refine the age of zonal boundaries or make the zonal subdivision of sediments more detailed.

The term "datum surface" was coined by Le Roy, who established the evolutionary appearance of *Orbulina universa*. He wrote: "...qualified points of lowest stratigraphic occurrence constitute a time datum to be termed the surface" of a certain species (Le Roy, 1948, p. 502). Therefore, a datum level or surface indicates the time point of evolutionary appearance or extinction of a certain species. Presently, when correlating deep-sea sections, researchers often estimate the age of a first and last occurrence levels of species in a certain section. This age can depend on facies peculiarities or climatic changes in the region. It is the consideration of these zonal or stratigraphically important datums in several climatic areas that forms the basis of this paper.

It may be suggested that FADs and LADs of many planktonic foraminiferal species should differ in various climatic belts. A number of papers actually provide data on the discrepancy between datums of some forms in different latitudes, and on their greater or lesser reliability (Hills and Thierstein, 1989; Weaver and Clement, 1986).

It seems likely that a thorough study of microfauna from oceanic sections, displaying high deposition rates and good paleomagnetic records, and located in different climatic areas, can help in solving the problem. This paper concerns the stratigraphy of Late Cenozoic deposits at one drilling site in the subtropical North Atlantic. The main objective was to estimate and discuss the FAD and LAD ages of planktonic foramin-

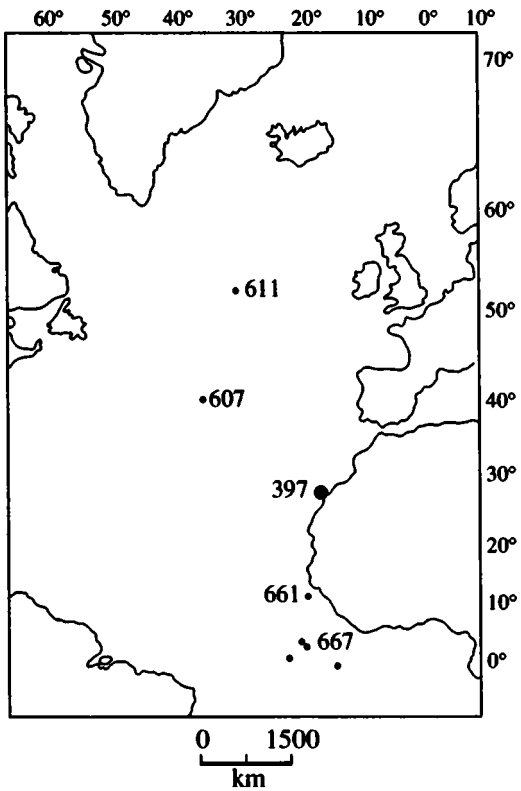

Fig. 1. Location of DSDP Site 397.

iferal zonal species and to test zonations applied in the studied region.

Site 397 (26°50.7′N, 15°10.8′W) is located nearby the northwestern coast of Africa at a depth of 2900 m (Fig. 1). There is a thick section of Upper Cenozoic deposits accumulated under high deposition rate. The available paleomagnetic record permits the estimation of FAD and LAD ages of planktonic foraminiferal zonal species. The samples were taken with intervals of 1.5 m which correspond to resolution of 0.02 m.y. Planktonic foraminifers do not bear evidence of dissolution, and their reworking was recorded only in Core 12 and in the lowermost Core 14. To facilitate comparisons with previous works, the age of datums were calculated using the accumulation rate values and the time scale of Berggren (Berggren *et al.*, 1985). The same ages calibrated against the more recent scale (Berggren *et al.*, 1995) are cited in the table. I acknowledge the DSDP leadership for donation of the core material.

PLANKTONIC FORAMINIFERAL DATUMS

Correlating FADs and LADs of planktonic foraminiferal zonal species at Site 397 with those from other climatic areas, I used information on corresponding datums from the equatorial and tropical North Atlantic at Sites 661–667, ODP Leg 108, 10°–2° N (Weaver and Raymo, 1989), and on Sites 606–611 of the DSDP Leg 94 from the middle latitudes (Weaver and Clement, 1986).

The first occurrence of *Globorotalia margaritae* is recorded in Site 397 within Chron 5 dated back to 5.64 Ma (Fig. 2) that is in good agreement with age dates for Sites 607–611 (Weaver and Clement, 1986) and coincides with the inference of Berggren who estimated it to be at 5.6 Ma (Berggren *et al.*, 1985). In this paper, the first finding of the early form *G. margaritae primitiva* in Sample 397-45-5, 32–36 cm, is considered as the appearance level of *G. margaritae*. Somewhat above, this form is followed by *G. margaritae margaritae* and then, in the same core, by *G. margaritae evoluta*. In the equatorial and tropical sites drilled by Leg 108, the appearance of *G. margaritae* varies from datum of 4.6–5.38 Ma at Site 661A to that of 5.7–6.2 Ma at Site 664D. It should be, however, noted that holes of Leg 108 do not all have the complete paleomagnetic control: the Matuyama/Gauss magnetic boundary was the lowest one clearly recognized in the holes (Weaver and Raymo, 1989). *G. margaritae* was widespread in most climatic belts of the North Atlantic. Even on the Rockall Plateau at Sites 403–406, 56°N, it is common enough to enable the distinction of the corresponding early Pliocene zone (Krasheninnikov and Bylinskaya, 1994). Judging from data on the nearby Site 611 with paleomagnetic record, *G. margaritae* appeared there 5.28–5.44 Ma ago (Weaver and Clement, 1986). Thus, this datum can be considered reliable in many areas of the North Atlantic. North of 56°N, *G. margaritae* sharply decreases in abundance, being missing in general from sediments. However, a single specimen of this form was recorded at the base of the Pliocene close to the inversion of coiling direction of *Neogloboquadrina atlantica* as far north as at Site 407, 63°56′ N (Krasheninnikov and Bylinskaya, 1994).

The first occurrence of *Globorotalia crassaformis* is the next biostratigraphic event upward the section. Forms of *G. crassaformis* classified as its subspecies are of great importance for the late Pliocene and Quaternary stratigraphy in the Atlantic, and they are given special attention in the paper. At Site 397, first specimens of *G. crassaformis crassaformis* are recorded within the Thvera event of the Gilbert Chron (4.62 Ma), i.e., considerably earlier than it was estimated before. In the equatorial–tropical transect of Leg 108, the FAD of *G. crassaformis* in the northern tropical area is recorded at 4.28–4.55 Ma (18°–21°N, Sites 657A and 659A), thus being closest to our datum. In the southward direction, this datum becomes younger (Fig. 3), having the estimated age of 3.8–4.18 Ma at Sites 660–664, except for Site 667A, where the corresponding datum is dated back to 4.12–4.56 Ma. In temperate latitudes, Sites 606–611, the FAD of *G. crassaformis* corresponding to 4.06–4.30 Ma, or on the average to 4.12–4.16 Ma, is almost identical with that of *G. puncticulata* (Weaver and Clement, 1986). In most sites of Leg 108, the simultaneous FADs of these species were also recorded, except for Sites 660A and 664D, where *G.*

Calculated ages of planktonic foraminiferal datums, DSDP Site 397 (FO—first occurrence of species in the section; LO—last occurrence of species in the section)

Planktonic foraminiferal datums	Age (Ma)	
	calculated using scale by Berggren *et al.*, 1985	calculated using scale by Berggren *et al.*, 1995
FO *Globigerina calida calida*	0.97	1.05
LO *Globorotalia crassaformis viola*	1.56	1.67
FO *Globorotalia truncatulinoides*	1.88	1.95
LO *Globigerinoides obliquus extremus*	2.01	2.08
LO *Globorotalia miocenica*	2.18	2.26
LO *Globorotalia pertenuis*	2.4	2.55
LO *Globoquadrina altispira*	2.91	3.03
FO *Sphaeroidinella dehiscens*	3.01	3.12
LO *Sphaeroidinellopsis seminulina*	3.02	3.13
LO *Globorotalia margaritae*	3.09	3.16
LO *Globoquadrina dehiscens*	3.24	3.4
FO *Globorotalia miocenica*	3.58	3.81
FO *Globorotalia crassaformis viola*	3.74	3.98
FO *Globorotalia exilis*	4.16	4.5
FO *Globorotalia crassaformis ronda* *Globorotalia crassaformis hessi*	4.16	4.5
LO *Globigerina nepenthes*	4.42	4.87
FO *Globorotalia puncticulata*	4.42	4.87
FO *Globorotalia crassafromis crassaformis*	4.62	5.05
FO *Globorotalia margaritae*	5.64	6.21

crassaformis first appears earlier and later than *G. puncticulata*, correspondingly. The FAD of *Globorotalia puncticulata* at Site 397 corresponds to 4.42 Ma. As for the northern latitudes (56°–63°N), Sites 403 and 408, *G. puncticulata* appeared earlier than *G. crassaformis*, whereas at Site 407 both forms are coeval (Krasheninnikov and Bylinskaya, 1994). Therefore, FADs of both species or at least of one of them, are likely to be diachronous in the Atlantic. It is remarkable that the earliest FAD is recorded in the subtropical area.

The stratigraphic distribution of *Globorotalia crassaformis* subspecies was almost never considered, except for papers of Bolli who distinguished them (Bolli and Premoli Silva, 1973; Bolli and Saunders, 1985). Next after *G. crassaformis crassaformis*, specimens of *G. crassaformis ronda* and *G. crassaformis hessi* were encountered upward in the section at Site 397. They first appeared within the Nunivak event of the Gilbert Chron of 4.16 Ma. Whereas *G. crassaformis ronda* was previously recorded in the Pliocene sediments, for instance, at Site 403 (Krasheninnikov and Bylinskaya, 1994), *G. crassaformis hessi* was originally described as the index species of the corresponding Quaternary subzone (Bolli and Premoli Silva, 1973), and according to our data on drilling sites in the Rockall Plateau and to the north, this form actually appears not earlier than in the middle Pleistocene. However, at the tropical Site 368 (17°30′N), this subspecies was found in the Pliocene *Globorotalia miocenica* Zone, and southward at Site 366A (5°40′N), it is recorded in the Late Pliocene *Globorotalia tosaensis* Zone (unpublished data). Therefore, *G. crassaformis hessi* likely appeared earlier in the subtropical Atlantic, because its FADs to the north and south of the subtropical belt is younger.

The *Globorotalia crassaformis viola* FAD at Site 397 is near the top of the Cochiti event of the Gilbert Chron (3.74 Ma). According to Bolli, *G. crassaformis viola* occurs in the Caribbean beginning from the latest early Pliocene (Bolli and Saunders, 1985). In our samples, this subspecies also occurs in the upper part of the *Globorotalia margaritae* Zone at the equatorial–tropical Site 366A and near 56°N at Site 403. It may be suggested that the first occurrence of *G. crassaformis viola* in the Atlantic marks the comparatively reliable and stratigraphically convenient datum.

The *Globigerina nepenthes* LAD at Site 397 corresponds to the time point of 4.42 Ma at the top of the Sidufjall event of the Gilbert Chron. This datum is obviously diachronous and varies in different holes. In

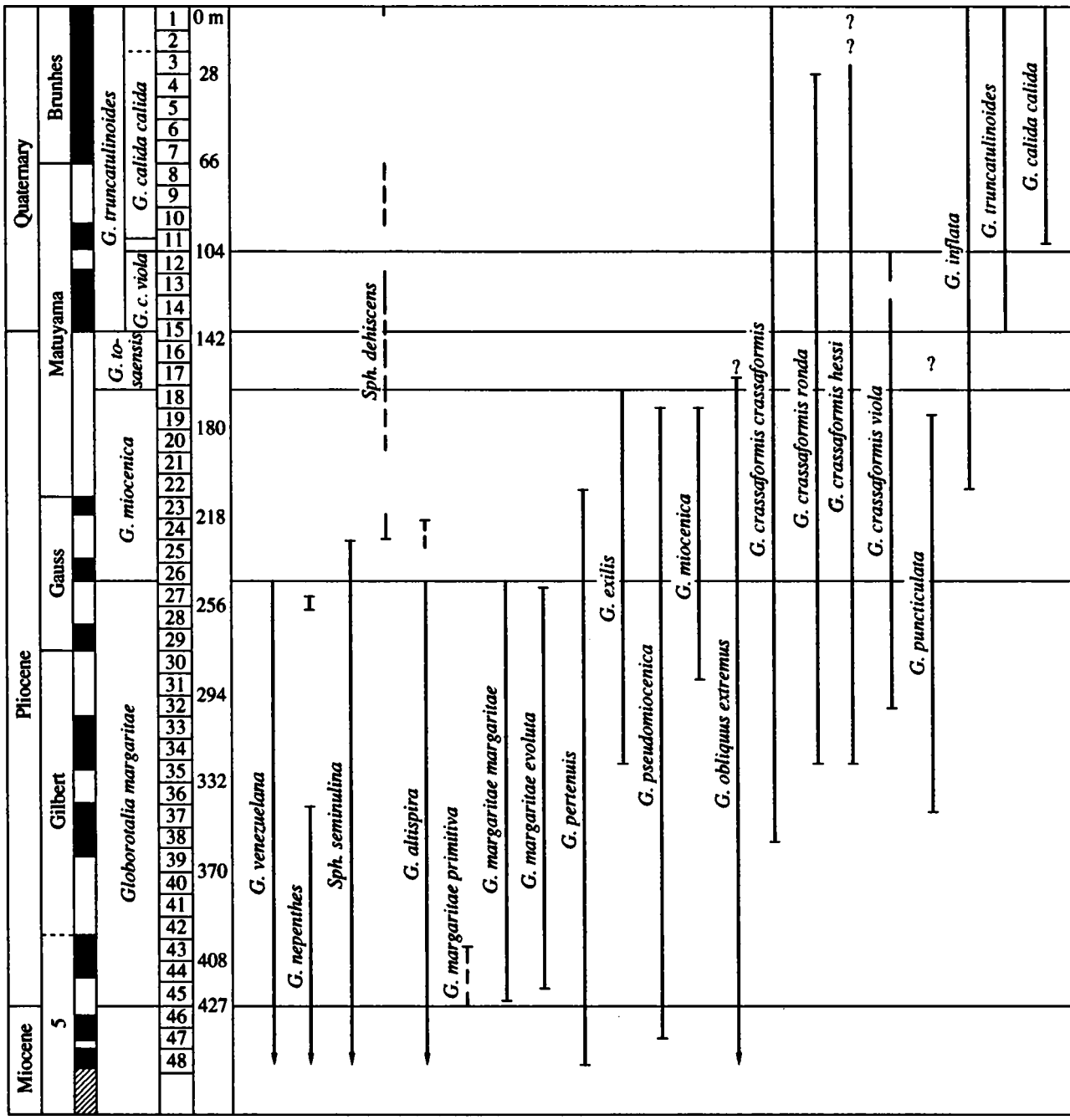

Fig. 2. Stratigraphic distribution of planktonic foraminifers in the Pliocene and Pleistocene deposits at Site 397.

the equatorial–tropical latitudes it ranges from 3.8 to 4.55 Ma (Weaver and Raymo, 1989). According to data of Leg 94, its estimated age varies from 4 Ma in Site 606, 37°N, to 4.7 Ma in Site 611 at 53°N (Weaver and Clement, 1986). In most publications, the *G. nepenthes* LAD is considered to be not younger than 3.8 Ma. However according to Bolli and Saunders (1985), *G. nepenthes* shows its last occurrence together with *Globorotalia margaritae* at the level of 3.2 Ma. It should be mentioned that at Site 397 last specimens of *G. nepenthes* are found at the time mark of 3.12 Ma in the uppermost *Globorotalia margaritae* Zone, but they are separated from our datum by a large interval, in which *G. nepenthes* is missing. Therefore, the last findings may represent the reworked specimens. The use of *G. nepenthes* in zonations or simply as the Early Pliocene guide fossil evidently requires verification of its extinction datum.

Other important stratigraphic events in the mid-early Pliocene are the FADs of *Globorotalia exilis* and

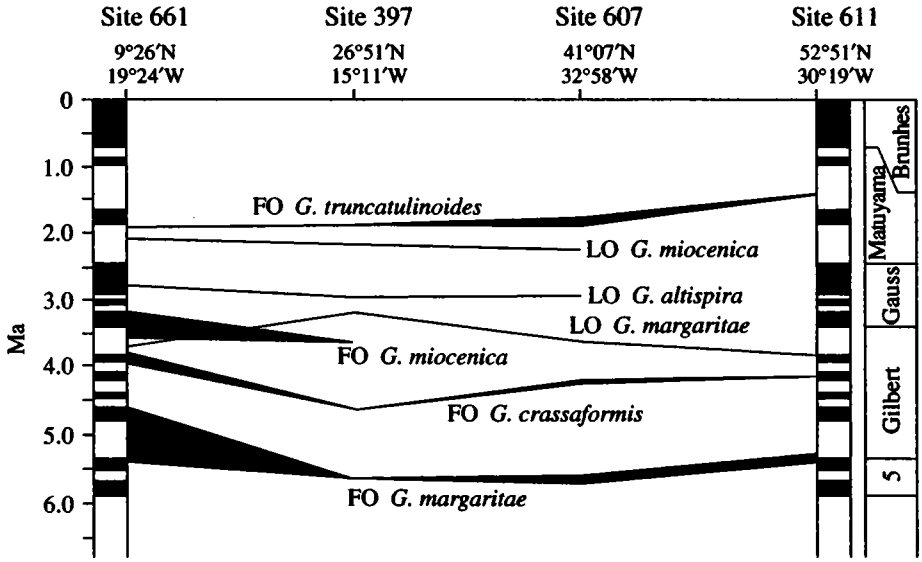

Fig. 3. First and last appearance datums of some Pliocene–Quaternary planktonic foraminifers in the North Atlantic (data on Site 661 are from Weaver & Raymo, 1989; those on Sites 607 and 611 from Weaver & Clement, 1986): FO—first occurrence of a species; LO—last occurrence of a species in the section.

Globorotalia miocenica. At Site 397, they are recorded at the levels of 4.16 and 3.58 Ma, correspondingly. In the Caribbean these species were also found in the uppermost *Globorotalia margaritae* Zone, though *G. miocenica* appeared earlier (Bolli and Saunders, 1985). The first occurrence of *G. miocenica* at Sites 657A and 659A, 18°–21°N, is close to our datum, being as old as 3.35–4.18 Ma (Weaver and Raymo, 1989).

Globoquadrina dehiscens occurs at Site 397 throughout the greater part of the *Globorotalia margaritae* Zone. Its LAD is equal to 3.24 Ma just below the base of the Mammoth event of the Gauss Chron. According to published data, in other sites this event occurred significantly earlier. Bolli detected the extinction of *G. dehiscens* in the lower half of the *Globorotalia margaritae* Zone, whose base coincides with the Miocene–Pliocene boundary (Bolli and Saunders, 1985). After Berggren (1977), the last occurrence of *G. dehiscens* marks the Miocene–Pliocene boundary at the base of the Gilbert Chron, i.e., at 5.35 Ma. At Sites 606–611, this species was not found in the upper Miocene deposits (Weaver and Clement, 1986). Finally, in the Leg 108 sites, LADs of this species vary greatly from 8.2–10.5 Ma at Site 659A to 4.6–5.38 Ma at Site 661A (Weaver and Raymo, 1989). The assumption that such a late occurrence of *G. dehiscens* at Site 397 resulted from redeposition is in my view unlikely, because the species occurs almost in every core, and other signs of redeposition are absent. Therefore, the *G. dehiscens* LAD appears to be insufficiently reliable for the use in zonal schemes.

The *Globorotalia margaritae* LAD is the next biostratigraphic event detected upward in the section. This datum is mentioned to be among the most reliable for the Pliocene planktonic foraminifers and has the estimated age of 3.66 Ma (Hills and Thierstein, 1989). However, my data on Site 397 show that *G. margaritae* shows its last occurrence at 3.09 Ma at the top of the Mammoth event. According to Bolli and Saunders (1985), the estimated age of this datum is 3.2 Ma. In the equatorial to tropical latitudes the estimated extinction age of *G. margaritae* varies from 3.3–3.5 Ma at Site 667 to 3.8–3.9 Ma at Site 660 (Weaver and Raymo, 1989). In the Leg 94 sites, this datum appears to be diachronous (Weaver and Clement, 1986). Even at closely spaced Sites 606 and 607, the last occurrence of *G. margaritae* is recorded at 3.5 and 3.7 Ma, correspondingly. Northward, this event took place even earlier: 4.13 Ma (Site 609 at 50°N) and 4.5 Ma (Site 610 at 53°N).

Sphaeroidinellopsis seminulina has its last occurrence in the deposits of Site 397 nearby the Kaena event of the Gauss Chron (3.02 Ma), i.e., close to that inferred by Hills and Thierstein (1989), who argued for the reliability of this datum. The authors noted that the extinction of *S. seminulina* in the Atlantic is recorded in a number of holes at 3.07 Ma. In the equatorial and tropical area, the estimated ages of this datum range from 2.8 to 3.2 Ma, with the mean value of 3.0 Ma (Weaver and Raymo, 1989). In the Mediterranean, it is also recorded within the Kaena event (Cita, 1975). After Weaver and Clement (1986), in temperate latitudes the *S. seminulina* LAD is almost identical at Sites 606 and 607: 3.0 and 3.04 Ma. The authors stated that north of 41°N this species is lacking, though, from our data, its single specimens occur in the lower half of the Pliocene up to the Rockall Plateau, namely, to 56°N (Krasheninnikov and Bylinskaya, 1994). Finally, Bolli and Saun-

ders (1985) also recorded the extinction of *S. seminulina* in the lower part of the *Globorotalia miocenica* Zone of the middle Pliocene. Thus, this event appears to be synchronous in the most part of the Atlantic and is actually reliable as the stratigraphic marker.

The *Sphaeroidinella dehiscens* FAD at Site 397 is almost coincident with the extinction of *Sphaeroidinellopsis seminulina* at 3.0 Ma, though, after Bolli and Saunders (1985), *S. dehiscens* appears earlier, in the upper part of the *Globorotalia margaritae* Zone.

Such an important biostratigraphic event as the last occurrence of *Globoquadrina altispira* is recorded at Site 397 near the top of the Kaena Subchron at 2.91 Ma. This age is in good agreement with most records in other drill holes. According to Hills and Thierstein (1989), it corresponds to 2.94 Ma and is one of the most reliable datums in the Pliocene. In the equatorial and tropical latitudes *G. altispira* shows its last occurrence between 2.85 and 3.05 Ma in all sites of Leg 108 (Weaver and Raymo, 1989). At Sites 606 and 607 of the southern temperate area, this LAD is recorded between 2.93 and 2.99 Ma (Weaver and Clement, 1986). During Leg 94, *G. altispira* was not found north of 41°N, although its single specimens occur in the Pliocene deposits up to 56°N (Krasheninnikov and Bylinskaya, 1994). Bolli and Saunders (1985) detected the extinction of *G. altispira* in the lower half of the *Globorotalia miocenica* Zone.

The *Globorotalia pertenuis* LAD at Site 397 was clearly recognized in the middle of the *Globorotalia miocenica* Zone at 2.4 Ma within a cool climatic interval.

Globorotalia miocenica has its last occurrence in the studied hole at 2.18 Ma that is in good agreement with data of Leg 94. For instance, at Sites 606 and 607 its LADs are at 2.2 and 2.3 Ma, respectively (Weaver and Clement, 1986). In the equatorial and tropical Atlantic, the species LAD ranges from 2.0 to 2.3 Ma (Weaver and Raymo, 1989). The same or close estimated ages of the datum, namely, 2.1 and 2.2 Ma, are reported by Mazzei *et al.* (1979) and Berggren *et al.* (1985), correspondingly. Therefore, the extinction level of *G. miocenica* is among reliable Pliocene datums in the Atlantic. The LADs of *G. pseudomiocenica* and *G. exilis* are very close to it. These three datums mark the upper boundary of the *Globorotalia miocenica* Zone and the base of the latest Pliocene *Globorotalia tosaensis* Zone.

The *Globigerinoides obliquus extremus* LAD at Site 397 is just below the Olduvai event at 2.0 Ma. This age is earlier than that for Sites 606–609 in temperate latitudes, where the estimated age of the *G. obliquus extremus* LAD ranges from 1.62 to 1.9 Ma (Weaver and Clement, 1986). In the equatorial and tropical area, this event is recorded at about 1.8 Ma (Weaver and Raymo, 1989). According to other records, the datum age varies greatly in different holes having a mean value of 2.43 Ma (Hills and Thierstein, 1989).

The *Globorotalia truncatulinoides* FAD is recorded at Site 397 at the base of the Olduvai Subchron (1.88 Ma). The mean estimated age of its first occurrence in some other holes is 1.99 Ma (Hills and Thierstein, 1989). In equatorial and tropical latitudes, this datum ranges from 1.8 to 2.0 Ma, excluding holes, where sediments evidently underwent dissolution (Weaver and Raymo, 1989). In the temperate area, *G. truncatulinoides* first appears at 1.84–1.85 Ma (Sites 606–609), and at 1.8 Ma (Site 610). However, at Site 611 situated somewhat west of the latter though at the same latitude of 53°N, this event is recorded much later, namely, at 1.35 Ma (Weaver and Clement, 1986). From data on Site 403 (56°N) northward of the Rockall Plateau, *G. truncatulinoides* first appears at about 1.25 Ma (Krasheninnikov and Bylinskaya, 1994). Thus, the *G. truncatulinoides* FAD appears to be sufficiently reliable in the North Atlantic from the equator to 50°N. In a northward direction it becomes younger.

The *Globorotalia crassaformis viola* LAD is distinguished for the first time at Site 397, where it is at 1.56 Ma. This event marks, in my view, the top of the lowest Pleistocene *Globorotalia crassaformis viola* Subzone. The estimated age of this datum is likely provisional, because at Site 397 there is a hiatus above the *G. crassaformis viola* LAD.

The first occurrence of *Globigerina calida calida* marks one more Quaternary biostratigraphic event. Its age was originally estimated to be 0.14 Ma (Bolli and Premoli Silva, 1973). However, after the later correlation with nannoplankton data, it became apparent that the datum is older. In the deep-sea cores from the tropical Atlantic, the *G. calida calida* FAD is recorded within the *Pseudoemiliania lacunosa* Zone at the base of the Brunhes Chron at 0.73 Ma (Bylinskaya and Golovina, 1990). In the temperate belt (the Rockall Plateau), the first occurrence of this form also roughly corresponds to the Matuyama/Brunhes paleomagnetic boundary (Krasheninnikov and Bylinskaya, 1994). At Site 397, *G. calida calida* first appears within the Jaramillo event at 0.97 Ma. The age of this datum, therefore, still needs further refinement.

STRATIGRAPHIC ZONAL SCHEMES

Among zonations suggested for the subtropical Atlantic, let us consider those of Bolli and Saunders (1985) and Berggren (1977). Bolli and Saunders distinguished three major zones with four subzones in the Pliocene, and one zone with five subzones in the Pleistocene (Fig. 4). Their subdivision of the Pliocene is based on such major evolutionary events as the last occurrences of *Globorotalia margaritae* and *G. miocenica*. Main Pleistocene zonal markers are the first occurrence levels of *G. truncatulinoides, G. crassaformis hessi, Globigerina calida calida*, and *Globorotalia fimbriata* (Bolli and Premoli Silva, 1973; Bolli and Saunders, 1985).

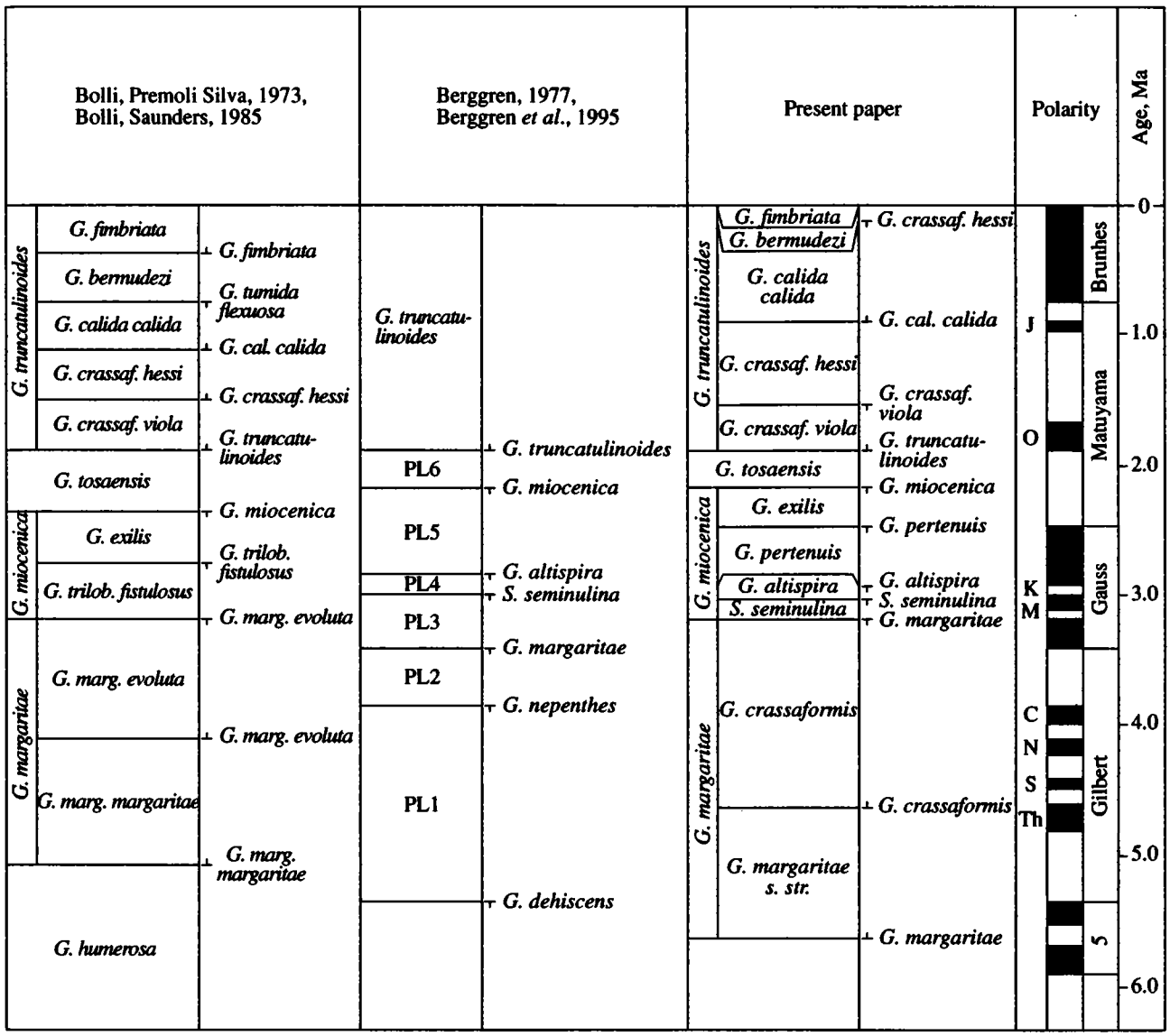

Fig. 4. Correlation of the Pliocene–Quaternary interval of planktonic foraminiferal zonations: (1) first occurrence of index species; (2) last occurrence of index species.

The zonation of Berggren (1977) includes six Pliocene zones, boundaries between which are defined by two listed events and by extinction levels of *Globigerina nepenthes, Sphaeroidinellopsis seminulina*, and *Globoguardrina altispira*. The Pleistocene in this zonation corresponds to the undivided *Globorotalia truncatulinoides* Zone.

Despite the use of mainly different index species and various number of subdivisions, the comparison of these zonations reveals their conceptual similarity (Fig. 4). Within the stratigraphic range of *Globorotalia margaritae*, Bolli distinguished subzones of two subspecies, whereas Berggren divided zones PL1 and PL2 by the boundary at the extinction level of *Globigerina nepenthes*. The next *Globorotalia miocenica* Zone after Bolli is subdivided into two subzones and corresponds in its range to zones PL3, PL4, and PL5 of Berggren. In the latter zonation, the boundaries correspond to extinction levels of *Sphaeroidinellopsis seminulina* and *Globoquadrina altispira*, correspondingly. Finally, the *Globorotalia tosaensis* Zone of Bolli corresponds in the assemblage composition to the zone PL6 of Berggren.

The comparison of these zonations with the established stratigraphy of Site 397 leads to the following. The recognition of the lower Pliocene *Globorotalia margaritae* Zone, the index species of which is well identified as abundant and widespread in the North Atlantic from the equator almost to 60°N, is by far warranted. It is divided into two parts in both discussed zonations. However, the boundary between the latter at the extinction level of *Globigerina nepenthes* (Berg-

gren, 1977) is not quite valid, as this event is clearly diachronous in different areas of the Atlantic, and, in addition, this species is rather scarce. The stratigraphic boundary within the *G. margaritae* Zone may be placed at the first occurrence level of *Globorotalia crassaformis* (Fig. 4) or *G. puncticulata*, as it is done in one of the proposed zonations (Weaver and Clement, 1986). It is precisely the occurrence and abundance of these species that give the characteristic appearance to all the overlying planktonic foraminiferal assemblages.

The recognition of the middle Pliocene *Globorotalia miocenica* Zone (Bolli and Saunders, 1985) is also essential, though it seems to be valid only up to a latitude of about 40°N because of the thermophilic character of the index species. The subdivision of this zone into smaller units also encounters some difficulties. According to Bolli, the boundary between *Globigerinoides trilobus fistulosus* and *Globorotalia exilis* subzones corresponds to the last occurrence of the former species and is valid only in the narrow equatorial and tropical gyre, for it is missing north of about 15°N. The *Sphaeroidinellopsis seminulina* and *Globoquadrina altispira* LADs used in the Berggren's zonation are more clear and reliable, though the corresponding zones PL3 and PL4 span each only about 0.11 m.y. The *Globorotalia pertenuis* LAD that occurs in this interval (the table) may be used as the stratigraphic boundary marker as well. Accordingly, I suggest to distinguish the middle Pliocene *Globorotalia miocenica* Zone with the *Sphaeroidinellopsis seminulina, Globoquadrina altispira, Globorotalia pertenuis* and *G. exilis* subzones (Fig. 4).

The late Pliocene *Globorotalia tosaensis* Zone is valid as the interval from the extinction of *G. miocenica* to the first occurrence of *G. truncatulinoides.*

Judging from the records at Site 397, the subdivision of Quaternary sediments into five subzones suggested by Bolli and Saunders (1985), is fully justified here like in many other Atlantic regions. In my opinion, only the definitions of these subzones should be somewhat modified. For instance, the upper boundary of the lowest subzone should be placed at the extinction level of the index form, whereas the top of the *Globigerina calida calida* Subzone should coincide with the last occurrence of *Globorotalia crassaformis hessi.*

Thus, I suggest the following modification of Bolli's zonation for the Pliocene–Quaternary interval. In the Early Pliocene, I distinguish the *Globorotalia margaritae* Zone, whose lower boundary corresponds to the first occurrence of the index species at 5.6 Ma, and the upper one to its extinction at about 3.1–3.2 Ma. The zone is subdivided into *Globorotalia margaritae* s. str. and *Globorotalia crassaformis* subzones with the boundary between them defined by the first occurrence of *G. crassaformis* at 4.6 Ma. In other climatic areas, this boundary is most likely at a higher stratigraphic level. In the Late Pliocene, two major units are distinguished: the *Globorotalia miocenica* Zone subdivided into the *Sphaeroidinellopsis seminulina, Globoquadrina altispira, Globorotalia pertenuis*, and *Globorotalia exilis* subzones with the boundaries placed at 3.02, 2.9, 2.4, and 2.2 Ma, respectively; and the *Globorotalia tosaensis* Zone with the upper boundary corresponding to the first occurrence of *Globorotalia truncatulinoides* at 1.88 Ma. Within the Pleistocene *Globorotalia truncatulinoides* Zone, the following five subzones are distinguished: (1) *Globorotalia crassaformis viola* Subzone ranging up to the extinction level of the index form at 1.56 Ma; (2) *Globorotalia crassaformis hessi* Subzone; (3) *Globigerina calida calida* Subzone ranging from the first occurrence of the index form at 0.9–0.73 Ma to the extinction level of *Globorotalia crassaformis hessi* approximately at 0.18–0.12 Ma; (4) *Globigerina bermudezi* Subzone; and (5) *Globorotalia fimbriata* Subzone corresponding to the Holocene and beginning from the first occurrence level of the index species.

CONCLUSION

(1) The stratigraphic study of the Site 397 section in the subtropical Atlantic and comparison of results with planktonic foraminiferal zonation in other climatic areas revealed in the Pliocene and Pleistocene of the Atlantic five reliable biostratigraphic events, namely, the first occurrence of *Globorotalia margaritae* and *G. crassaformis viola*, and the extinction of *Sphaeroidinellopsis seminulina, Globoquadrina altispira*, and *Globorotalia miocenica.* The contemporaneous first occurrence of *G. truncatulinoides* in the North Atlantic between the equator and 50°N is also proved. The following datums may be assumed reliable, which, however, calls for further investigation: those of the first occurrence of *Globorotalia crassaformis ronda, G. exilis, Globigerina calida calida*, and those of the extinction of *Globorotalia pertenuis* and *G. crassaformis viola.*

(2) The Pliocene–Quaternary sediments of the subtropical Atlantic were subdivided in good accordance with the zonation of Bolli and Saunders (1985), though slightly modified. The deposits of the upper Pliocene part of the section are subdivided using a combination of Berggren (1977) and Bolli's zonations with a minor modification (see Fig. 4). It should be noted that the suggested variant of zonation may be applied, on our evidence, in the North Atlantic from the equator to 40°N. Further to the north, some biozones of most thermophilic species may be missing from the late Pliocene part of zonation. However, the general structure of zonation remains unchanged up to the latitude of about 60°N.

(3) The correlation of planktonic foraminiferal FADs and LADs in different climatic areas (Fig. 3) showed that a number of species appeared in the subtropical Atlantic earlier than in more northern and southern regions. These species are *Globorotalia crassaformis crassaformis, G. crassaformis hessi, G.*

puncticulata, G. miocenica, and, likely, *Globigerina calida calida*. The detailed study of nannoplankton from the same site revealed that typical large *Gephyrocapsa oceanica* first appeared there 1.78 Ma ago (L.A. Golovina, personal communication), i.e., earlier than its datum estimated before. All these results favor the hypothesis suggesting a subtropical radiation center of the species, however, the problem as to whether they support the hypothesis, or are a coincidence, is still open.

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Reference Section of the Log Krasnyi Loess and Soil Formation of the Lower Pleistocene in Upper Reaches of the Don River

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Abstract—Materials of geological, paleontological (small mammals and freshwater mollusks), paleopedological, palynological, paleomagnetic and magnetic studies are considered to characterize the Log Krasnyi loess-and-soil formation of the lower Pleistocene in the upper reaches of the Don River. The comprehensive investigation revealed four thermochrons in the lower Pleistocene sequence. The upper three thermochrons are within the Brunhes Chron. The second thermochron from the top and partly the underlying cryochron span the interval of geomagnetic excursion. Alluvium corresponding in time to the Matuyama Chron and underlying the loess-and-soil deposits yielded the most representative collection of the early Tiraspol' theriofauna that was studied in detail. The Log Krasnyi section can be considered as a stratotype of the loess-and-soil formation in the upper reaches of the Don River.

Key words: *stratigraphy, microtheriofauna, palynology, thermochron, excursus*

INTRODUCTION

The upper Don River basin is an area most suitable for the climatostratigraphic study of the Pliocene–lower Pleistocene deposits, because the sequence of alluvial beds of a complicated structure and a series of subaerial formations were deposited in the valley and water divide areas, respectively at that time. Another favorable point is the presence of regional stratigraphic marker such as the Don Moraine left by the glacier tongue. Many microtheriological data indicate the early Pleistocene age of the moraine (Breslav *et al.*, 1992; Velichko *et al.*, 1992; Krasnenkov *et al.*, 1977, 1980; Udartsov, 1985). Some of the studied sections showed that the Don Moraine and respective erosional products are wedged into the complex of alluvial deposits and subaerial formations. This is favorable for "linking" the stratigraphic sequences (Krasnenkov *et al.*, 1977, 1984). In the upper Don River basin, an assemblage of small early Pleistocene mammals was collected and studied in detail. In abundance and diversity of species, this collection is the largest one in the glacial zone of Eastern Europe and characterizes fauna of the earliest periods to the final stages of the early Pleistocene (Krasnenkov *et al.*, 1984). In 1994, an expedition under the leadership of R.V. Krasnenkov carried out in the paleopedological, palynological and paleomagnetic studies in the Log Krasnyi area.

The present paper is dedicated to the analysis of geological, microtheriological, paleopedological, paleomagnetic, and palynological data from the Log Krasnyi section in order to establish a stratotype of the loess-and-soil formation in the upper reaches of the Don River. The Log Krasnyi section is one of the most complete stratigraphic sections in the area. It is represented by a series of loess and soil beds dated back to the late Eopleistocene and the early Pleistocene (predating the Don Morain). The glacial deposits and subaerial formations are exposed in local branches of the ravine that opens into the valley from the Izmaila River, a small tributary of the Don River.

GEOLOGICAL DESCRIPTION OF THE SECTION (AFTER KRASNENKOV)

The Log Krasnyi sequence is located in a ravine of the same name, 2.5 km to the west-northwest from the northern outskirts of Uryv Village (Pokrovka) in the Ostrogozhsk district of the Voronezh region (Fig. 1).

A series of strippings exhibits the following beds below the recent soil (Figs. 2, 4).

(1) QI dns: Don Moraine bed of brownish unlaminated loam enclosing pebbles and rare small boulders

† Deceased.

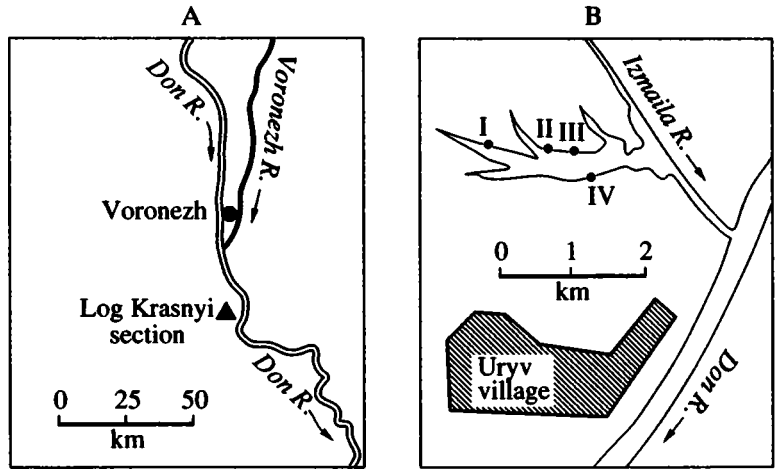

Fig. 1. The Log Krasnyi locality (A) and position of studied strippings I–IV (B).

of chalk, rare inclusions of violet quartzites of the Shoksha type, and boulders of granites, shales and veined quartz (1.3 m).

(2) QI dns: Don Moraine bed of dark gray unlaminated clay bearing rare inclusions, mostly of chalk; debris rocks transported from distant areas are rare; the lower bed boundary is smooth.

(3) pdQI vrsh: Vol'naya Vershina buried soil; the bed is composed of dark gray humified and aggregated loam fragmented and showing traces of dynamic action of the glacier in the upper part; a mass of loam veins penetrates deep into the underlying bed (1.3 m).

(4) prQI rost: Rostush Loess; the pale yellow loesslike bed unlaminated but disturbed by veins from Bed 3; veins of the bed deeply penetrate into Bed 5; assemblage of mollusks consists of *Vallanyia teneulabris* (al. Br.), *Pupilla sterri* Voith., and *P. muscorum* (L) identified by Krasnenkov (0.4 m).

(5) pdQI trn: Ternovka buried soil; the humified loam bed aggregated and getting darker in the lower part, particularly in "shaggs" that penetrate into Bed 6; the lower part is also disturbed by mole burrows (1.3 m).

(6) prQI klsn: Koleshnino Loess; the brownish to greenish loesslike bed broken at the top by veinlets and "shaggs" descending from Bed 5; lime concretions are concentrated along fissures (1.3 m).

(7) pdQI trz: Troitskoe buried soil; the bed of dark gray, slightly brownish loam humified and intensively aggregated; the bed shows cyclicity of soil formation with the phase boundaries marked by lime veinlets and small concretions; the lower cycle of soil formation is most spectacular (3.15 m).

(8) prQI: light-colored loam considerably reworked by processes of soil formation; the bed likely represents BC horizon of the Troitskoe soil (0.2 m).

(9) pdQI zr: Zor'kino buried soil; structurized, brown to reddish loam with rare inclusions of lime concretions (1.5 m).

(10) aQI pk: Pokrovka Horizon; intercalation of reddish–brown fine-grained sand and greenish sandy loam; the bed contains crumbs of chalk (1.9 m).

(11) aQI pp: Petropavlovka Horizon; horizontally bedded light-colored sand with clay intercalations and chalk pebbles at the base; a considerable collection of small mammals (see the table) is extracted from the lower part of the bed (2.0 m).

(12) K_2t: white chalk bed with a destroyed surface.

PALEOPEDOLOGICAL CHARACTERISTICS OF THE SECTION

The paleopedological analysis of the loess and soil beds in the Log Krasnyi section is carried out on the basis of field observations. The conclusions are tentative, because the paleomorphological situation at the site is studied insufficiently, and soil horizons in separate outcrops are correlated mainly on the basis of paleomagnetic data.

At the main Exposure I just below the Don Moraine, the buried soil is attributed by Krasnenkov to the Vol'naya Vershina unit (nomenclature of this and other soil beds is after Krasnenkov, 1992). The buried soil experienced diagenetic transformations, and its upper part is apparently missing. The A1–BC soil profile is 1.4-m thick. The main stage of its formation was humification, and additional study is necessary to verify conclusion about the soil type. The bed rock of the soil is the Rostush Loess, the horizon of which overlies with a hiatus the Ternovka Soil.

The Ternovka Soil is the second buried soil unit below the Don Morain; it shows the A1-Bca-C profile and is about 2-m thick being formed under the grass vegetation. During the first stages of its formation, the humidity regime was contrast and resulted in appearance of specific ferruginous concretions. Later on, these concretions were covered with carbonate film,

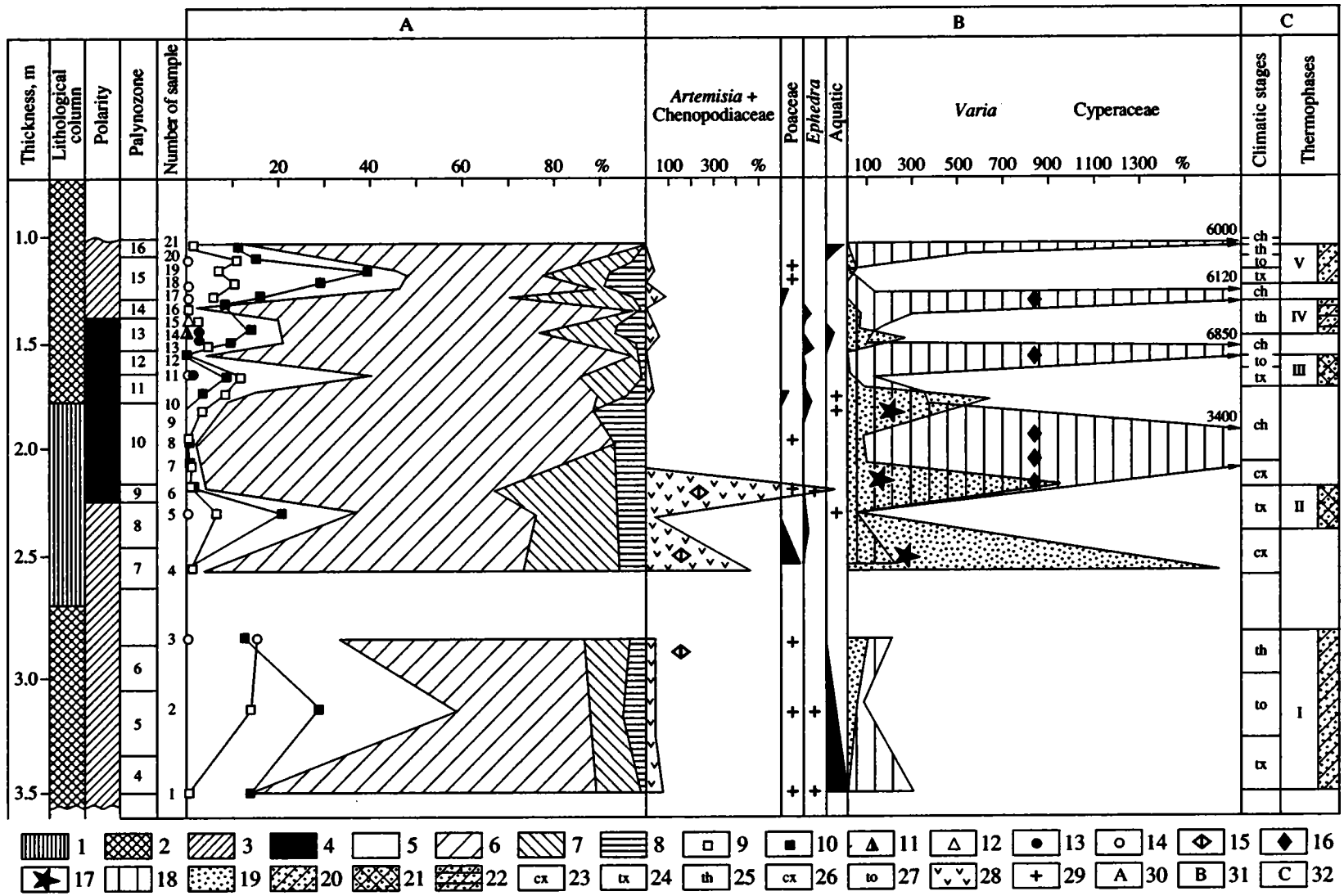

Fig. 2. Spore–pollen diagram for the Troitskoe and Ternovka soils and Koleshnino Loess, Exposure II (spore–pollen diagram for all rock varieties of the exposure see in Pospelova *et al.*, 1998): (1) loess; (2) fossil soil; (3) normal polarity; (4) reversed polarity; (5) pollen of arboreal species; (6) pollen of mesophilic herbaceous-frutescent plants; (7) pollen of xerophilic herbaceous-frutescent plants; (8) spores; (9) pollen of alder; (10) total amount of broad-leaved arboreal species; (11) pollen of fir; (12) pollen of spruce; (13) pollen of pine; (14) pollen of birch; (15) dominance levels of Chenopodiaceae; (16) dominance levels of herbaceous groups; (17) dominance levels of Cyperaceae; (18) *Varia*; (19) Cyperaceae; (20) Interglacial; (21) Interstadial; (22) Interphasial; (23–27) climatic stages: (23) cryoxeric, (24) thermoxeric, (25) thermohygrotic, (26) cryohygrotic, (27) thermal optimum; (28) total of *Artemisia* and Chenopodiaceae pollen; (29) levels with single pollen grains; (30) summarized spore–pollen diagram (percentage is calculated from the total sum of pollen and spores in the sample); (31) content of herbaceous-frutescent plants calculated as percentage of arboreal pollen; (32) climatic stages of thermophases.

and their coexistence with rather loose carbonate concretions indicates that the carbonate illuviation was one of the soil formation factors. These features probably implies that the soil was formed in the meadow environment in the first half of the warm epoch, whereas later, when the steppe conditions developed, the soil acquired properties of the Chernozem type. Final stages of the soil formation were characterized by polygonal fissuring indicative of the temperature and humidity gradients and at the time of transition to the next stage of loess accumulation.

Below the Koleshnino Loess representing the bed rock for the overlying soil, there is a pedocomplex of two partly superimposed soils, which we correlate with the Troitskoe Soil. The uppermost soil bed shows the $A1_I$–$A1_{II}$–Bca profile and is about 3-m thick. The bed was formed in warm conditions below the grass vegetation. It seems to be polygenetic, because its humus layer is not homogeneous in color and can be divided into two subzones, the upper of which is less reworked by soil-forming processes. All these features indicate that the soil formation was accompanied by sedimentation. It is noteworthy that the Chernozem-like soil was formed above two-layer deposits, and the older soil layer was more differentiated in structure and served as the local water–impermeable horizon responsible for a slight gleying in the middle part of the profile.

The lowermost soil of this pedocomplex (Troitskoe Soil) represents the topless profile about 1 m thick in Exposure I. This soil greatly differs from all younger soil layers. Apparently, its profile is of the eluvial–illuvial type and has been formed in a sufficiently humid climate. The exposure shows only the lower part of the Btf horizon, whose material fills fissures, and the Bca horizon transitional to the loess bed. The Btf horizon is reddish brown, rich in iron. It shows micropolygonal fissures being very distinct in the exposure as the reference horizon. The Bca horizon contains abundant and large, very dense ("stonelike") carbonate concretions, whose genetic relations with the lower or upper soil layers of the pedocomplex are problematic. They could be also formed later under the effect of ground waters rich in calcium carbonate.

In Exposure II, located 250 m apart from the first one, a part of the loess–soil sequence is washed out, because there is the inclined erosion contact between these layers and the Don Morain. Stratigraphic correlation between soil layers observable in Exposures I and II is not unambiguous on the basis of paleopedological data. Apparently, in this place, the Vol'naya Vershina Soil, Rostush Loess and a part of the Ternovka Soil were eroded by the glacier.

Under the moraine, the exposed lower part of the Chernozem-like soil may correspond to the Ternovka Soil as it is evident from geomagnetic excursion (PMZ-2) recorded at this level. During the general stage of its formation, this soil experienced mainly humification without traces of illuviation of clay, iron, or carbonates.

Small mammal fauna of the Log Krasnyi section

Insectivora	Number of specimens
Desmana sp.	5
Drepanosorex sp.	1
Sorex sp.	1
Rodentia	
Citellus sp.	20
Allactaga sp.	1
Allocricetus sp.	3
Mimomys pusillus Mehely	101
M. ex gr. *intermedius* Newton	10
M. cf. *hintoni* Fejar	1
M. reidi Hinton	2
Villanyia cf. *hangaricus*	3
Clethrionomys aff. *sokolovi* Topachevski	7
Allophaiomys–Pitymys	4
Pitymys hintony Kretzoi	18
P. gregaloides Hinton	1
P. cf. *duodecimeostatus* Selys–Longchanps	4
Pitymys sp.	1
Microtus oeconomus Pallas	2
M. cf. *hyperboreus* Vinogradov	1
Microtus sp.	98
Prolagurus pannonicus Kormos	14
Prolagurus sp.	18
Eolagurus simplicidens	2
Gliridae gen.	1

In the course of formation of the soil, the landscape was of the steppe, rather than of the forest–steppe type. At the final formation stage, the most evident process of micropolygonal fissuring was probably associated with the grown temperature and humidity gradients. The loess material was accumulated later and filled numerous small fissures in the soil.

The pedocomplex of Exposure I correlative with the Troitskoe Soil lies within the 1-m interval under the Ternovka Soil. Its structure is somewhat different. The brownish humus horizon 1.3 m thick is undifferentiated into subhorizons, but becomes darker with depth. It grades downward into the Bt horizon 1.5 m thick, where fissures are filled with dark gray humic material of a different color than that of the A1 horizon of the exposure. In the lower part, the pedocomplex has a horizon of accumulated dense carbonate concretions surrounded by calcareous material. The soil is of the heavy loam type and overlies reddish to white sandy-

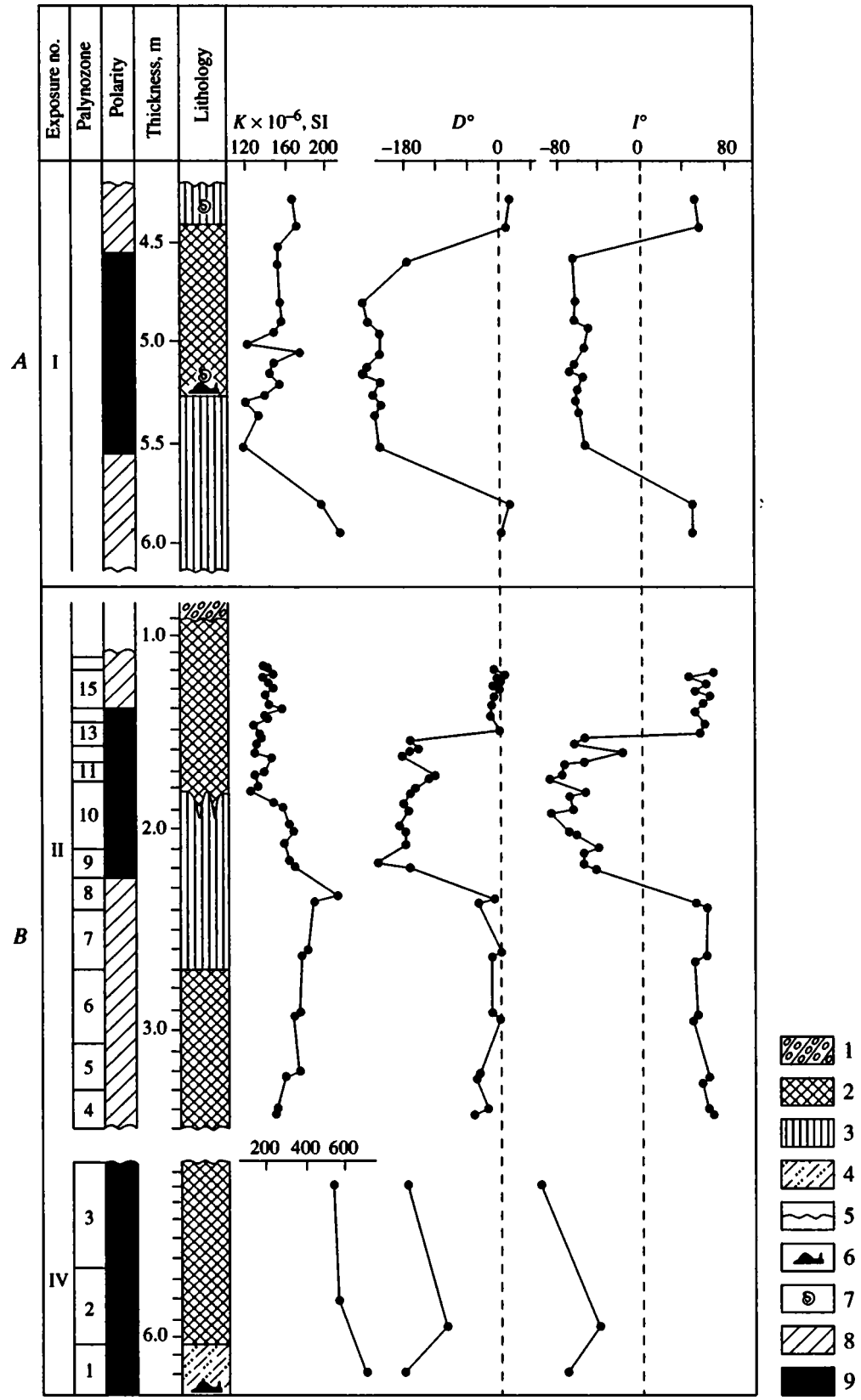

Fig. 3. Paleomagnetic and palynological characteristics of the Log Krasnyi sections: (1) moraine loam; (2) buried soil; (3) loess; (4) alluvium; (5) hiatus; (6) fauna; (7) freshwater mollusks; (8) normal paleomagnetic zone; (9) reversed paleomagnetic zone.

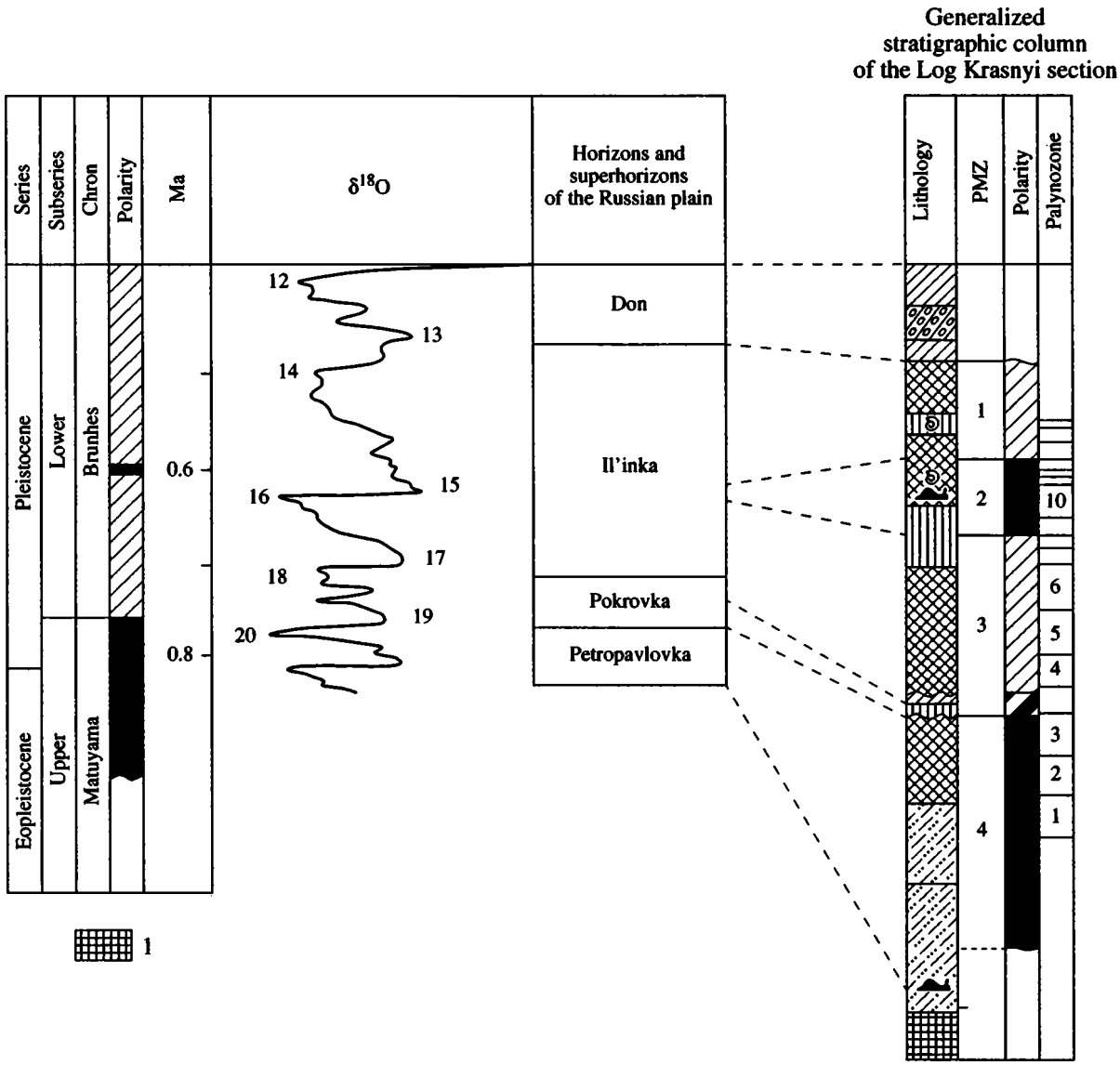

Fig. 4. Log Krasnyi loess–soil section correlated with the stratigraphic sequence of the Russian plain and oxygen-isotope variation curve (after Krasnenkov, 1992). (1) Gretaceous. For other symbols see Fig. 3.

loam alluvium entirely corresponding to the reversed polarity zone.

All fossil soils in Exposures III and IV are thicker and show weak differentiation into genetic horizons. In Exposure IV, only one thick soil resting on reddish alluvium is distinguished. This structure of the sections implies that the beds can be classed with the slope deposits. Without additional paleogeomorphological data, these soils in a series of exposures can be correlated only on the basis of paleomagnetic data. According to Krasnenkov, this level corresponds to the Zor'kino Soil of the Eopleistocene thus being a part of the Petropavlovka Horizon of the ISC scheme (*Reshеniya* ..., 1986). Soil at this level is oldest of all soils studied in Exposures I and II.

FAUNA OF SMALL MAMMALS

At the base of the Ternovka Soil, Krasnenkov identified the remains of small mammals, among them the teeth of zokors, and also of terrestrial mollusks. Since zokor remains were found in the middle of the Il'inka Horizon (there were no finds below this level), the middle Il'inka (middle Tiraspol') age of the Ternovka Soil is inferred.

The collection of small mammals was obtained by Krasnenkov and studied by N.E. Kazantseva with consultations by A.K. Agadzhanyan. The micromammal assemblage is very diverse, about three times greater than in the well-known collection recovered from the stratotype section near the Petropavlovka village. Rhizodont voles of the *Mimomys* genus are represented by

four species with a considerable dominance of *M. pusillus* Villania. Among three bush voles, the archaic *Pitymys hintony* is dominant. Another characteristic feature is the high abundance of gray voles (*Microtus*), two species of which are most archaic elements of the Tiraspol' fauna. The lagurides are represented only by Pannonian lemmings. From the ecological viewpoint, the list of small mammals given in table implies the forest-steppe landscape, because lemmings, including yellow steppe forms, and jerboas coexisted with abundant redbacked and bush voles. Dormouse was also present amid them as a typical forest element.

The analyzed collection of small mammals characterizes for the first time the Petropavlovka fauna from the glacial zone of the East European platform. The microtheriofauna under consideration is attributable to the early Tiraspol' assemblage of mammals. The fauna suggests the early Pleistocene age of loess and soil beds below the Don Morain.

PALYNOLOGY

Palynology of the Log Krasnyi section is studied in 24 rock samples (52 macerations). The spore–pollen diagram reveals 16 palynozones, and the same number of the vegetation and climate phases is reconstructed (Pospelova *et al.*, 1998). The red alluvium beneath the loess-and-soil beds contains the Tiraspol' assemblage of small mammals and characterizes the epoch of dominance of arboreal vegetation, namely of the polydominant alder–oak–elm forests. This optimum corresponds to the thermophase. The red-brown Zor'kino Soil was formed during the thermal phases 2 and 3 and is rich in flora consisting of 46 taxa, a part of which does not grow now in the study area. The overlying gray Troitskoe Soil initially accumulated in the cryophase 4 and afterward during phases 5 and 6. During the climatic optimum (phase 5), the hornbeam forests flourished in the area. The warm and very humid climate of that time is indicated by finds of single pollen grains of *Pterocarya* in combination with dominant pollen of hornbeam (Fig. 2).

The Koleshnino Loess and overlying Ternovka Soil were palynologically studied in greater detail (34 macerations). The Koleshnino Loess horizon was formed in an almost forestfree landscape, as all 13 macerations, except three that correspond to the interstadial (interglacial phase 8), contain less than 10% of arboreal pollen. The paleogeographic situation at the accumulation time of the Koleshnino loess was, however, unstable. The period of loess formation is marked by consecutive changes of three xerotic (7–9) and one hygrotic (10) floral successions indicative of climatic fluctuations (Fig. 2). There are two peculiarities in the development of floral successions of phases 8 and 9, the boundary between which coincides with the lower boundary of the Log Krasnyi geomagnetic excursion. First, the transition from the thermal optimum (phase 8) to the cryoxerotic succession (phase 9) is very sudden. The absence of gradual changes in the types of reconstructed vegetation is usually attributed to hiatuses. Inside the Koleshnino Loess, this feature is not recorded. Second, in our case there is a certain inversion in the normal, though different in water supply, stages of the glacial–interglacial cycle, because we inferred a succession of cold–humid, cold–dry, warm–humid, and warm–dry stages. A similar succession of stages was revealed as early as the 1960s by Grichuk and Grichuk (1960).

At present, many researchers associate epochs of loess formation with cryogenic stages of the glacial cycle. Palynologists, however, often obtain different paleogeographical reconstructions for these epochs. For example, according to N.S. Bolikhovskaya, "the accumulation of loess horizons in the East European province of the loess–soil formation (LSF) took place during all stages of glacial climatic rythms, including the interstadial and interphase warmings, and likewise the xerothermal stages and endothermal coolings of the interglacial climatic rythms ... Only the cryoxeric stages of glaciation on the Russian plain were characterized by the extreme localization of the soil cover proper and by the widespread deposition of loess beds." (Bolikhovskaya, 1995; p. 239).

The Ternovka Soil was formed during five climatic phases corresponding to palynozones 11–15 (Fig. 2). The inferred successions show that this soil was formed during three forest–steppe phases divided by short but sharp and almost forestless phases (12 and 14), when widespread in the area were stunted herbage groupings. Similar regularities in vegetation development are recorded in the Strelitsa quarry section of the upper Don River basin (Bolikhovskaya, 1995).

PALEOMAGNETIC DATA

The reconnaissance paleomagnetic studies of the loess–soil beds in the Log Krasnyi section revealed the reversed polarity of rocks of the Petropavlovka Formation (the Matuyama Chron). The anomalous direction of magnetization in rocks was recorded at the base of Ternovka soil and in the underlying Koleshnino Loess (Semenov, 1994).

Rock samples for paleomagnetic studies were collected in the interval from the Rostush Loess and to the Petropavlovka Alluvium, inclusive. The continuous sampling was applied to the second fossil (Ternovka) soil and to the upper part of the underlying Koleshnino Loess.

The magnetic susceptibility of rocks (K) was measured *in situ* and using the rock cubes at the laboratory. The K-parameters vary along the section from 120–200×10^{-6} SI (the Ternovka and Troitskoe soils and Koleshnino loess between them) to 500–700×10^{-6} SI (the Zor'kino Soil underlying with a hiatus the Troitskoe soil and alluvium below the former). Values of remanent magnetization (Jn) follow, on the whole, variations of K-parameters. In order to determine the

direction of initial magnetization, rock samples were heated stepwise up to 600°C, having the geomagnetic field compensated to 5γ, and the Jn component analysis of rocks was also carried out (Pospelova *et al.*, 1997).

Four paleomagnetic zones (PMZ) were identified as a result of the Jn component analysis in the Log Krasnyi section (Fig. 3). The Vol'naya Vershina Soil, underlying Rostush Loess, and an upper part of the Ternovka Soil correspond to PMZ-1 of normal magnetization. The remaining part of the Ternovka Soil and the upper part of the Koleshnino Loess are within PMZ-2 of reversed magnetization. The rest of the Koleshnino Loess and the Troitskoe Soil turn out to be in PMZ-3 of normal polarity. The PMZ-4 of reversed polarity spans the alluvium and partly the red-brown Zor'kino Soil of the Petropavlovka Horizon, the lower part of which yielded ancient representatives of the Tiraspol' fauna of small mammals.

The geological and microtheriological data are used to compare the PMZ-4 with the Matuyama Chron. The overlying loess-and-soil sequence (PMZ-3, PMZ-2, PMZ-1) was formed in the first half of the Brunhes Chron. The thin (about 1 m) PMZ-2 of reversed polarity is reliably determined in three exposures of the Log Krasnyi section and corresponds to the Log Krasnyi excursion termed after the place of discovery. Another interpretation, according to which PMZ-2 is attributed to the end of the Matuyama Chron, contradicts data on the evolution of small mammals (Agadzhanyan, 1992).

Factor Q=Jn/0.5K at 400°C and Jn/Jrs ratios were used to appraise the geomagnetic field intensity in the formation period of rock samples from Exposures I–IV. According to these parameters, the geomagnetic field intensity during the excursion was 3–4 times less than the stationary field intensity before and after this event. Within the Log Krasnyi excursion, the geomagnetic field is of the reversed polarity. The calculated virtual geomagnetic poles (VGP) are located around the southern geographic pole within the circle of 68°S. The average VGP coordinates are $\varphi = 82°S$, $\lambda = 2°E$ (Pospelova *et al.*, 1997).

By studying the temperature dependence of the saturation magnetization Js versus the remanent saturation magnetization Jrs, the Curie points and blocking temperatures of magnetic minerals in the rocks were determined. The composition of magnetic minerals in the studied rocks was found to be different. The main carrier of magnetization is the fine-grained hematite in the case of the Ternovka and Troitskoe gray soils, Koleshnino loess, and light-colored Petropavlovka alluvium, all sampled at the Exposure IV. In the Zor'kino Soil and underlying red alluvium, the carrier is magnetite.

The rocks differ not only in the composition of magnetic minerals, but also in size and concentration of magnetic grains. The studied rocks are divided into two groups by the size of magnetic grains. The first group with smaller grains comprises samples of the Petropavlovka alluvium, Troitskoe Soil, and Koleshnino Loess. Samples from the Ternovka Soil and underlying loess represent the second group and have larger grains of magnetic minerals. The highest concentration of magnetic grains is detected in the Koleshnino Loess and Troitskoe Soil; the lowest concentration is in the Ternovka Soil. The nearer is the location of a given sample to the moraine, the lower is the magnetic grains concentration and the larger is their size. It can be presumed that the small magnetic grains in rocks underlying the Don Moraine were brought by the glacier. The difference in the composition of magnetic minerals from the studied rocks implies either the different origin of magnetic grains, or their different provenances (Pospelova *et al.*, 1997).

The comparison of magnetic and palynological data did not reveal a direct connection between the scalar magnetic parameters of rocks and climatic changes in the period of their accumulation. This seems to be a consequence of different composition and size range of magnetic minerals in the rocks of the studied section.

However, a comparison of paleomagnetic and palynological data for the same samples showed a certain connection between geomagnetic and climatic changes. During the Log Krasnyi excursion, climatic environments were changeable. The commencement of an excursion was in the long period of cooling (palynozones 9 and 10), and during its second half, the cooling was succeeded by two climatic optimums (palynozones II and I3) divided by a short period of cooling (palynozone 12). It was assumed that, as a rule, the geomagnetic excursions are characteristic of warm climatic periods (Petrova *et al.*, 1992). During the time of the Log Krasnyi excursion, both the sudden cooling and warming of climate took place. The beginning and end of the Log Krasnyi geomagnetic excursion were confined to sudden climatic coolings. After the geomagnetic excursion, the cooling was followed by a distinct interglacial period (palynozone 15), whereas the optimum before the onset of the excursion corresponded to the interglacial or interstadial palynozone 8. The revealed fluctuations of geomagnetic field and paleoclimate appear to be interrelated.

DISCUSSION

The representative collection of small mammals and freshwater mollusks together with palynological and paleopedological data on the Log Krasnyi section reveal at least four thermochrons in the lower Pleistocene deposits below the Don Moraine. Three of them span the stratigraphic interval from the Don to Petropavlovka Horizon thus corresponding to the Brunhes Chron.

The correlation of Quaternary climatic stages in upper reaches of the Don River with the oxygen-isotope variation curve (Bassinot *et al.*, 1994) is of considerable interest. The most problematic is the correlation of the Don glaciation with one of the stages of that

curve. Some geologists (e.g., Shik, 1992) refer this cryochron to the 16th stage. Krasnenkov (1992), while considering geological and paleontological data on the advanced evolution of small mammals during this stratigraphic interval, correlated the Don cryochron with the 12th stage of the oxygen-isotope scale. According to data obtained by Agadzhanyan (1992), four subassemblages of the Tiraspol' fauna succeeded one another, and this discredits their correlation with the time period of 120–130 ka spanning stages 17–19. The results of the absolute age determination of the Don Moraine (400–493 Ma according to TL, RTL, OSL) do not contradict this statement. Therefore, it seems reasonable to correlate the Petropavlovka, Troitskoe, Ternovka and Vol'naya Vershina thermochrons with stages 19, 17, 15 and 13, respectively (Fig.4).

In the Log Krasnyi section, at the level of Ternovka soil and in the upper part of the underlying loess, paleomagnetic studies revealed an interval of reversed magnetization that is interpreted as the geomagnetic excursion in the lower part of the Brunhes Chron. The excursion coincides with the end of the 15th and the beginning of the 16th oxygen-isotope stage. Earlier, in the Odintsovo stratotype area (the Akulovo bore-hole section), the geomagnetic excursion was recorded in the Akulovo interglacial deposits below the Don Moraine (Semenov, 1983). This excursus apparently is an analog of the Log Krasnyi excursion. According to certain data (Harland *et al.*, 1990; Pospelova, 1989), reliable geomagnetic excursions in the Brunhes chron have not been established below the 16th stage. This implies that the Akulovo Beds in the Moscow region are coeval with the Ternovka thermochron in the climatostratigraphic scheme for the upper Don River basin.

Results of this research open a new perspective to correlate the Lower Pleistocene stratigraphic scheme of the upper Don River basin, which is now the most detailed and substantiated for this age interval in Europe (Turner, 1996), with stratigraphic schemes of most important stratotype areas of the Odintsovo and Roslavl' glacial–lacustrine deposits. This is a new step toward the solution of key problems in the correlation of the lower Pleistocene deposits of the Russian plain.

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The Manzurka Alluvium of the Baikal Region: Palynology, Stratigraphy, Accumulation Stages

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Abstract—The Manzurka Alluvium fills ancient valleys of the Baikal–Lena watershed. Previously, the sequence of the constrative alluvium was considered to be of the Eopleistocene age. Latest investigations showed that despite the lithological uniformity, the sequence includes several stratigraphic units ranging in age from the Pliocene to late Pleistocene. The Manzurka Alluvium is a typical example of the polychronous sedimentary complex of the intracontinental frost-free areas that were formed under gradual environmental changes.

Key words: *Manzurka Alluvium, polychronous complex, Neogene–Quaternary deposits, ancient valleys, Western Baikal region.*

INTRODUCTION

The importance of the investigation of deposits in ancient valleys in the western Baikal region was first stressed by A.L. Chekanovskii more than 120 years ago. Pavlovskii and Frolova (1941) were first to describe fragments of ancient valleys in this region. First comprehensive studies of deposits in ancient valleys were headed by N.A. Logachev (Logachev and Abramova, 1958; Logachev *et al.*, 1964). The deposits were united into the Manzurka Formation (sequence) of the Eopleistocene age. Subsequent studies brought an additional information on the structural and distribution patterns of the Manzurka sequence (Adamenko *et al.*, 1971, 1980; Logachev *et al.*, 1974; Zamaraev *et al.*, 1976; *Stratigrafiya SSSR ...*, 1984; Kononov, 1988; Osadchii, 1987). The recent studies of these deposits using the radiothermoluminescence (RTL) method of dating allowed a conclusion that, despite the lithological uniformity, they were formed during different periods (*Manzurskii allyuvii*, 1995).

DISTRIBUTION OF THE MANZURKA ALLUVIUM

The Manzurka Alluvium fills the ancient valley stretching from the lower courses of the Bol'shaya Goloustnaya River near the Lake Baikal coast (Fig. 1) northeastward, along the piedmonts of the Primorskii Ridge. Then, the valley deviates to the north and passes along the Bugul'deika, Ungura, and Manzurka rivers crossing marginal morphological structures of the Siberian platform and enters the Lena River valley in the vicinity of the Kachug Settlement. The ancient valley has several branches along the Khodontsa and Anga rivers, where it opens toward Lake Baikal. Another ancient valley that is located near the Bugul'deika River mouth and separated from Baikal by a narrow crest stretching in parallel to the shore line occupies a particular position. Its wide bottom is at a relative height exceeding 100 m. One more branch of the ancient valley along the Anga River flowing into the Lena River near the Kachug Settlement is filled with the Anga Alluvium (Logachev *et al.*, 1964; Adamenko *et al.*, 1980). In the Cenozoic paleogeographic reconstruction for the southern part of East Siberia, the ancient valleys filled with the Manzurka alluvium are considered as representing the past watercourses from the Lake Baikal into the Lena River basin that ceased after the formation of the Angara River source area.

STRATIGRAPHY OF DEPOSITS FILLING ANCIENT VALLEYS OF THE WESTERN BAIKAL REGION

Pliocene Horizons of the Manzurka Alluvium

The oldest deposits are recovered in the Northern and Alagui sections in the Bugul'deika River basin. In the latter, the 10-m-thick sequence consists of sandy–pebbly material with whitish sandy and gravelly interbeds. Sediments are characterized by the diagonal and cross bedding. The lower part of the sequence includes large boulders and angular blocks. The upper part 3 m thick is composed of Pleistocene alluvial deposits.

† Deceased.

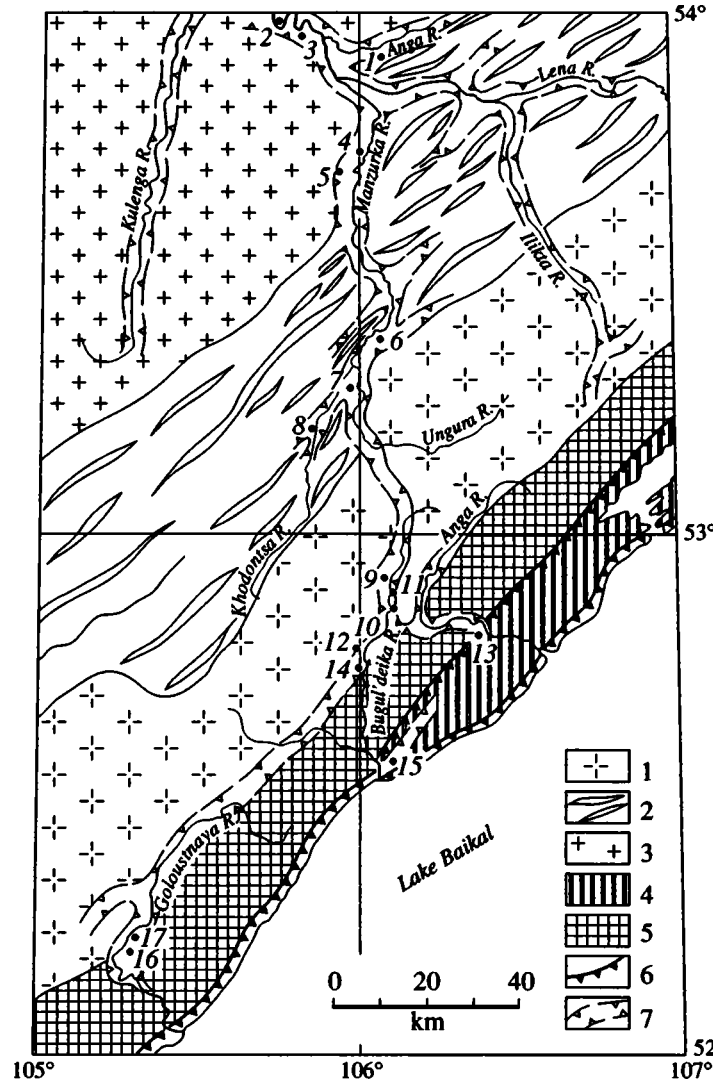

Fig. 1. Morphotectonic scheme of the interfluve area between the Lake Baikal and Lena River and localities of the studied Manzurka and Anga alluvium sections. (1–3) Siberian platform with its marginal elevated part (1) and a system of piedmont folds in the Baikal Trough (2), shown amid which are anticlinal heights and the Verkhnyaya Lena elevated scarp (3); (4, 5) Baikal rift zone with its marginal scarps and interbasinal bridges of the rift valley bottom (4) and inclined horsts at the western limb (5); (6) main tectonic scarps; (7) ancient valleys (after Logachev *et al.*, 1964 with modifications). Solid circles designate studied sections: Anga (*1*), Kisteneveo (*2*), Kachug (*3*), Kharbatovo (*4*), Nikelei (*5*), Kaptsygai (*6*), Podtok (*7*), Khogot (*8*), Ogul (*9*), Southern (*10*), Northern (*11*), Alagui (*12*), Elantsy (*13*), Khidusa (*14*), Bugul'deika (*15*), Pereval'nyi (*16*), Ilga (*17*).

Samples collected in the interval of 3–13 m yielded 26 preponderant genera of arboreal and frutescent vegetation. Abundant pollen of dark coniferous (*Tsuga, Picea, Abies, Pinus sibirica*) and pollen characteristic of mixed broad-leaved forests suggest the past existence of mountainous polydominant forest formations. Representatives of diverse (at the generic and species level) broad-leaved groups could be distributed in forests of a different type.

In the Northern section, alternating sandy–gravelly and sandy–gravelly–pebbly layers with distinct cross bedding occur up to a depth level of 15 m. The rocks are characterized by the development of Fe oxides in the form of spots, pours, bands, or complete ferrugination. The basal layer (15.0–15.8 m) is composed of compact variegated ferruginous loam with the Pliocene palynoassemblage.

The arboreal vegetation is mainly represented by pollen of *Pinus, Ulmus*, and dark coniferous taxa (*Tsuga* + *Abies* + *Picea* + *Pinus* s/g *Haploxylon*). Species indicative of composition of zoned forest formations are similar in both this and previous sections.

The early Manzurka pollen assemblages include four geographic groups of genera: (1) Panholarctic (*Abies, Picea, Larix, Pinus, Salix, Betula, Alnus, Euonumys, Sambucus, Lonicera, Myrica*); (2) American–

Eurasian (*Carpinus, Corylus, Ulmus, Acer, Tilia, Fraxinus, Ilex*); (3) American–Mediterranean–Asian (*Pterocarya, Juglans, Ostrya*); and (4) American–East Asian (*Tsuga, Diervilla*) genera.

Pollen assemblages from sections in the Bugul'deika River basin are similar to floras from the Pliocene deposits of mountainous areas of Northern Mongolia and eastern Transbaikal region: to those from the Member B of the Khirgis-Nur Formation in the Great Lakes Basin (Malaeva, 1987; Devyatkin *et al.*, 1989), from the alluvium underlying basalts in the Chulutu River valley in Khangai (Logachev *et al.*, 1982), and from the Onon Formation of the *whitish sequence* of the eastern Transbaikal region (Malaeva and Ufimtsev, 1976; Malaeva and Kramarenko, 1981; Simonov *et al.*, 1988). The composition of fossil floras indicates the past development of coniferous, coniferous–broad-leaved, and mixed forests in the southern mountainous areas of East Siberia. Pollen assemblages reflect a vertical zoning of the vegetable cover, which can be classed with the coniferous–broad-leaved forest vegetation of the nemoral type.

The stratigraphic position of the lower Manzurka deposits in the Bugul'deika River basin can be determined from comparison of their pollen assemblages with the above-mentioned fossil floras from neighboring regions, where additional stratigraphic criteria are known. The Khirgis-Nur Formation in the stratotype section Khirgis-Nur-2 is dated back to the early–middle Pliocene based on faunal assemblages (Devyatkin, 1981). Radiological datings of basalts overlying the Pliocene alluvium in the Chulutu River valley (ranging from 3.8 ± 0.2 to 4.0 ± 0.25 Ma) revealed the upper age limit for identical Pliocene pollen assemblages (Logachev *et al.*, 1982). The lower Manzurka deposits of the early–middle Pliocene age are united into the *Bugul'deika Horizon.*

A younger pollen assemblage of the Pliocene age was recovered in the Manzurka alluvium within the interfluve area of the Manzurka River and its left tributary, the Khal'skaya Ravine (the Kharbatovo section). The altitudinal difference between the watershed ridge and the Manzurka River bed is 80–85 m. This area hosts a group of quarries located along the distance of 2.5 km. In the northern quarry, samples were collected from the lower part of the exposed sequence 20–25 m thick. The sampled section is that of the lenticular shingle layers with lenses of white sand. Sediments are ferruginous; pebbles are well-rounded and covered by ferrugination films. Pebbles are mainly represented by volcanics, granites (often weathered), quartzites, and sandstones. In the southern quarry, only a part of the ancient alluvium sequence composed of sand and gravel is exposed. Here, inequigranular sands and gravel with interbeds of small-pebble and sandy material show the cross-lenticular to diagonal bedding and underlie loose deposits. The thickness of the sequence is about 8 m.

The pollen assemblage from the Kharbatovo section includes 21 genera of arboreal vegetation. Preponderant is pollen of *Pinus sylvestris* and *Betula* sec. Albae. Dark-coniferous formations are poorly represented in pollen spectra (Fig. 2). Composition of the diverse broad-leaved plants is characteristic of valley communities consisting of *Alnus, Ulmus, Yuglans, Hippophae, Salix* and of slope (zoned) mixed broad-leaved forests of *Fagus, Quercus, Carpinus, Acer, Tilia*, and *Corylus.* Pollen of the herbaceous–fruticose group is dominated by representatives of families and genera that are of a wide ecological range. They include taxa characteristic of mainly meadow–forest (*Pyrola, Ranunculus, Rumex, Heracleum, Cacalia hastata*), meadow–steppe (*Pheum, Pulsatilla, Galium, Trifolium*), and rocky steppe (*Polygonum angustifolium, Convolvulus, Eremurus, Ephedra, Thymus, Centaurea, Pulsatilla bungeana, Euphorbia discolor*) formations. The find of *Eremurus* (familiy Liliaceae) pollen is of particular importance for age determination. Modern representatives of the *Eremurus* genus are widespread in the Mediterranean–Asian region. In the time period under consideration, the western part of the genus distribution area reached the Baikal region. The connection of the considered flora with the South European–Asian floral province is confirmed by the presence of *Polystichum lobatulum* spores typical of the latter. The current distribution area of the last taxon is mainly confined to mountainous southern Europe and the Mediterranean. This species is characteristic of deciduous forests with *Carpinus* in the Caucasus and with *Fagus* in the Carpathians (Grichuk and Monoszon, 1971). Thus, the Kharbatovo flora contains some elements (*Fagus, Eremurus, Polystichum lobatum*) indicating its geographic connection with the Mediterranean floral province and thus suggesting the ancient age of the flora. In the Manzurka River valley, the old alluvium of the Kharbatovo Horizon occupies the highest (85–90 m) level of accumulation.

Younger layers of the Manzurka alluvium are exposed in the Podtok section (Logachev *et al.*, 1964; Zamaraev *et al.*, 1976; Adamenko *et al.*, 1980). The upper part of the section is composed of shingle cemented by whitish clayey sand. Pebbles are mainly represented by quartz and, to a lesser extent, by quartzites, dolomites, jasperoids, chalcedony, and siliceous rocks (interval of 0–1 m). These sediments are underlain by shingle cemented by ocherous–brown loam grading into ocher-colored to whitish gray clayey sand with pebble (interval of 1–2.4 m). Below, there are alternating strata of whitish clayey sand with pebble and boulders, brownish ocherous clayey sand, and loam with pebble, all of them filling incisions in a river bed. The altitudinal difference between the base of the quarry and the Ungura River bed is 42–43 m; the total thickness of loose deposits does not exceed 7–9 m. The arboreal pollen is dominated by *Pinus sylvestris, Betula*, and less common *Ulmus, Corylus, Tilia*, and *Yugalnas.* The herbaceous group is represented by pollen of

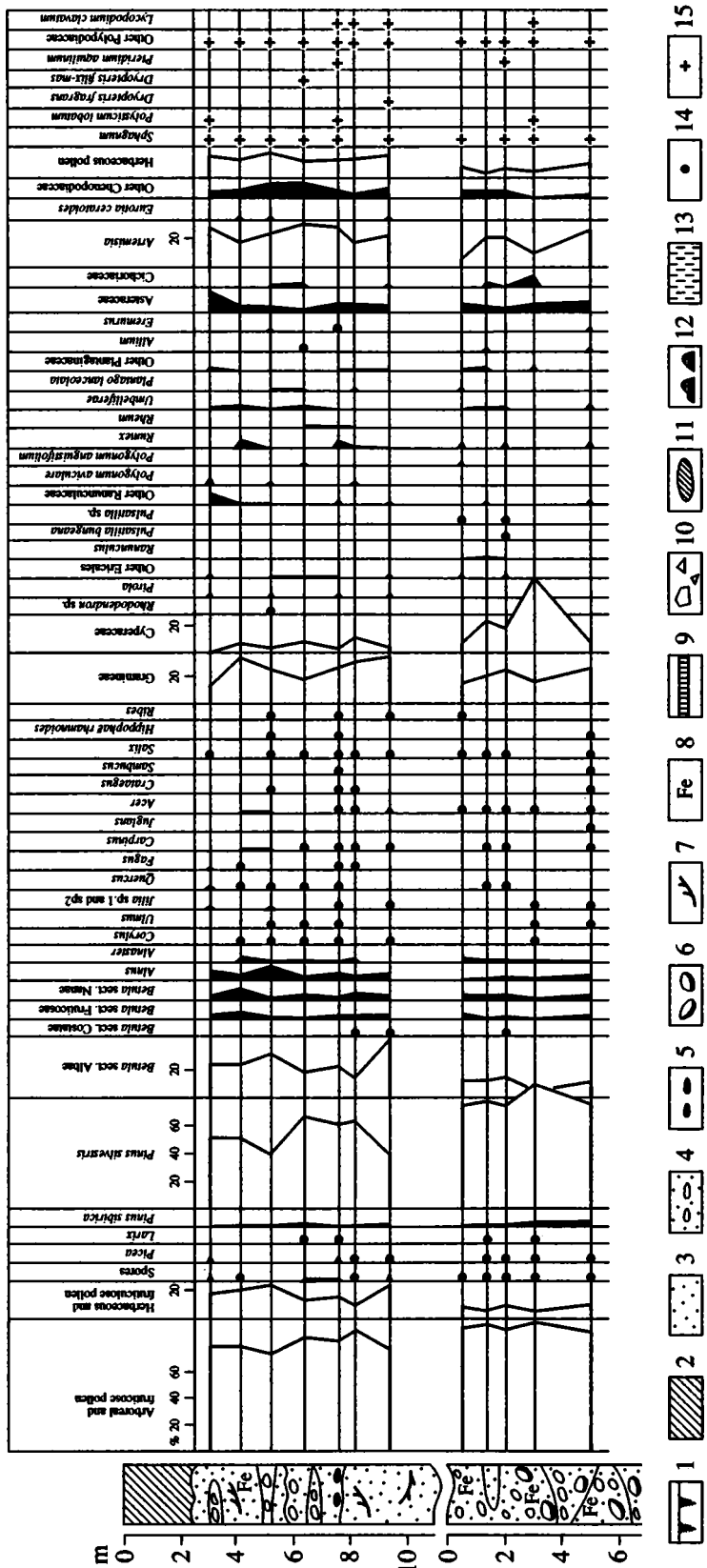

Fig. 2. Spore–pollen diagram for Pliocene N_2^2 deposits of the Kharbatovo section, the Manzurka River valley: (1) soil–vegetation layer; (2) blanket deposits; (3) sand; (4) gravel and pebble; (5) clayey rolls; (6) boulders; (7) bedding; (8) ferrugination; (9) loam; (10) shingle; (11) blocks of bed rocks of the Angara Group; (12) "nests" of ferruginous loam; (13) slope deposits; (14) pollen and spore abundance less than 1%; (15) rare pollen and spore grains.

Gramineae, *Artemisia*, and Cyperaceae. They are accompanied by pollen of widespread representatives from families Chenopodiaceae, Plantaginaceae, Unbeliferae, Cruciferae, Asteraceae, Cichoriaceae, and Urticaceae, and from genera *Allium, Ranunculus, Rumex, Polygonum, and Potentilla*. The pollen assemblage corresponds to the light-coniferous and mixed forest formations with an admixture of broad-leaved plants.

The considered deposits are united into the *Podtok Horizon* of the Manzurka alluvium (N_2^2).

The Pliocene Horizon of the Anga Alluvium

The Anga alluvium is distributed in the Anga River valley, the right-hand tributary of the Lena River (Logachev *et al.*, 1964). It is studied in the left bank of the Anga River, 14 km away of the Kachug Settlement.

The Anga malacofauna is characteristic of facies of shallow-water river pools and oxbow-lakes, as it is evident from presence of abundant *Corbicula, Pisidium amnicum*, and *Valvata*. Remarkable is a mixing of mollusks from different facies as a result of their redeposition. New materials obtained add information on ecological properties of the complex (Popova 1968, 1981) but do not change basically the available concepts.

The malacofauna is represented by 47 species, 14 of which are first recorded in the unit under discussion. Eight taxa missing from current assemblages of the Baikal region are *Valvata pulchella, Lymnaea stagnalis angensis,* Planorbis corneus, Planorbis sieversi, P. starobogatovi,* Helicorbis minima,* Corbicula praebaicalensis,** and *C. ferghanensis*. Taxa noted by the asterisk became extinct. *Valvata pulchella* and *Planorbis corneus* dwell now in Europe and West Siberia. *Planorbis sieversi* is of a wide distribution in Asia. The *Planorbis* genus, which includes the extinct species *P. starobogatovi*, is lacking now in the Baikal region, and its distribution area extends from the west only to the Yenisei River. The presence of genus *Corbicula* is remarkable, because this taxon is widespread southward of 45°N in the west occurring in the northern Black Sea and Caspian Sea regions and in Central Asia, whereas in the east (along the sea coast in the Primor'e region), its representatives occur at somewhat higher latitudes. *C. ferghanensis*, one of the most abundant species in the Anga sequence, is described also from the Fergana depression. The malacofauna suggests that the average annual air temperature in the Baikal region at the Anga time was higher than +5°C, probably amounting to 10°C (the modern value is about –2°C).

In addition to elements of the Siberian subregion of the Paleoarctic region (Starobogatov, 1986), the Anga sequence contains some forms characteristic of the European–Central Asian subregion of the Paleoarctic region (*Valvata pulchella, Planorbarius corneus, Planorbis* spp.). A significant share of extinct species and forms, which migrated to the more southerly distribution area, the high average annual temperatures, and zoogeographical structural patterns suggest the late Pliocene (Akchagylian) age of the Anga sequence. Initially, it was considered as the late Eopleistocene–Tiraspolian in age (Logachev and Popova, 1962). Later, when molluscan collections of their own and of N.A. Yasamanov, O.M. Adamenko, and R.C Adamenko were better studied (Popova, 1981), the cited authors shared an opinion based on the analysis of small mammal assemblages (Adamenko, 1975), that the Anga sequence is middle Eopleistocene in age (schemes by Gromov *et al.*, 1965). In later schemes (Nikiforova and Aleksandrova, 1987), the sequence is dated back to the Pliocene–early Eopleistocene.

Recent additional data indicate the late Pliocene age of the Anga sequence. In the Anga alluvium recovered in quarries Rykovo-1 and Rykovo-2, S.M. Popova and I.V. Shibanova sampled *Juga* aff. *orchonica* and *J.* aff. *starobogatovi* almost identical to species described from the gray-colored detrital sequence of the Orkhon River in Mongolia (Popova, 1978). The radiological age of beds with *Juga orchonica, J. starobogatovi*, and mammal remains within the gray-colored sequence is $<3.1 \pm 0.2$ Ma (the date is obtained for basalts underlying brown- to gray-colored member), whereas, according to the paleontological evidence (Devyatkin, 1981), the beds correspond in age to the Pliocene–early Eopleistocene in the scheme by Gromov *et al.*, (1960). Inasmuch as the Rykovo sequence is younger than the Anga sequence, the above-mentioned fauna confirms the late Pliocene age of the latter.

The pollen assemblage from the Anga alluvium was sampled in the interval of 3.7–6.85 m. The arboreal vegetation is represented by 21 genera. Preponderant is pollen of *Betula* and *Pinus sylvestris*, and a significant part of the spectra is represented by pollen of dark coniferous forms: *Picea* (to 18–25%) and *Pinus sibirica* (to 20–26%). The group of exotic forms includes *Picea* sec. *Omorica, Corylus, Ulmus, Tilia, Carpinus, Quercus, Fagus, Acer*, and *Elaeagnus* species. The herbaceous assemblage is dominated by pollen of widespread families and genera with the pollen admixture of aqueous vegetation (*Typha, Myriophyllum, Nuphar, Polygonum amphibium*).

The pollen spectrum corresponds to vegetation of the forest belt. Forests in the ancient river basin included two main formations: coniferous–broad-leaved and mixed broad-leaved with several species of broad-leaved trees, the south boreal species included. Thermophilic forms were represented by *Fagus* (most thermophilic), *Carpinus*, and *Elaeaganus*. The average temperature of the coldest months estimated for two European *Fagus* species is – 4–8°C.

The comparison of the Anga and Kharbatovo palynofloras reveals the following. The pollen composition and abundance ratios of taxa in the former indicate the significant role of dark coniferous forests in the vegetation cover, which preserved relics of the early Manzurka flora (*Picea* sec. *Omorica*). The composition

of other forest-forming species in two considered floras was similar (at the generic level), although the share of broad-leaved forms was higher during the period of the Anga alluvium formation. Based on taxonomic composition and percentage of main components in the fossil assemblages, it can be inferred that the Kharabatovo flora bears some features inherited from the Anga flora. Changes in the ratio of arboreal to herbaceous pollen in response to some aridization are observed in the upper part of the Anga deposits. This level is also marked by the increased proportion of *Betula* pollen and by appearance of *Elaeaganus* pollen characteristic of the tugai formations in the forest–steppe and steppe zones. In the Kharbatovo flora, the influence of aridization is more distinct, as the content of dark coniferous pollen is minimal in pollen spectra and its diversity is low. The share of pollen of broad-leaved forms decreased here to 12%, whereas *Eremurus*, an edificator of steppe formations, first appeared, and *Hippopha rhamnoides* characteristic of valley vegetation in the steppe and forest–steppe zones is also present.

Analysis of two floras indicates that the Kharbatovo flora directly succeeds the Anga floral assemblage, because it bears some features inherited from the latter. Deposits recovered in the Kharbatovo quarries are defined as a stratotype of the *Khabartovo Horizon* of the Manzurka alluvium of the upper Pliocene (N_2^2).

Pleistocene Horizons of the Manzurka Alluvium

The *middle Pleistocene* deposits were studied in valleys of the Khodontsa (the Khogot section), Anga (the Elantsy section), Goloystnaya, and Bugul'deika rivers and in the interfluve area of the Bugul'deika and Kuyada rivers (the Bugul'deika section).

In the Khodontsa River valley, 9 km to the south of the Khogot Settlement, the section of the accumulative level 35–40 m high was studied. The palynospectra are dominated by arboreal and fruticose pollen mainly representing *Pinus sylvestris* and *Betula*. Pollen of *Ulmus, Tilia, Carpinus, Quercus, Acer*, and *Yuglans* is also present. The date of 390 ± 80 Ka was obtained using the RTL method (determinations were performed by O.A. Kulikov in the laboratory of radiation monitoring of environments and RTL-dating at the Moscow State University).

In the northern outskirts of the Elantsy Settlement, a fragment of the Manzurka alluvium is exposed at the level of 70–80 m above the waterline of the Anga River. The section is mainly composed of pebbly–gravelly and sandy–gravelly strongly ferruginous deposits with small lenses of gray loam. Rock fragments are mostly represented by quartz, quartzites, schists, granitoids, and gneisses. The RTL date for the rocks is 290 ± 58 Ka. Palynospectra are dominated by arboreal pollen (18 genera) with *Pinus sylvestris* and *Betula* being most abundant (Fig. 3, A). The herbaceous pollen is characteristic of meadow–forest and meadow–steppe vegetation (gramineous–herbaceous groupings).

Similar palynospectra are recorded in deposits occurring near the valley side in the interfluve area between the Bugul'deika and Kuyada rivers. The Bugul'deika section is located at a height of about 120 m above the Baikal surface level. The whitish sandy–pebbly–gravelly sequence is strongly ferruginous here. Rock fragments are represented by quartz, quartzites, and granitoids; many of them are weathered. Deposits are often cross-bedded. The arboreal–fruticose palynospectra are dominated by pollen of *Pinus sylvestris* and *Betula* (Fig. 3, B). The herbaceous pollen spectrum is characterized by a mixed composition of meadow–forest and meadow–steppe plant communities.

Thus, the ancient alluvium of the Elantsy and Bugul'deika sections yielded pollen assemblages of the forest and forest–steppe vegetation. Mixed pine–birch forests prevailed. Representatives of the broad-leaved flora were constituents of zoned (*Quercus, Tilia, Acer, Corylus*) and valley (*Alnus, Ulmus, Yuglans, Carpinus*) forests.

The middle Pleistocene deposits were also studied in the Goloystnaya River basin. The Pereval'nyi and Ilga sections (RTL date of 230 ± 60 Ka) are located in the local watershed area between the Ilga and Sukhaya rivers, the left-hand tributaries of the Goloustnaya River. Palynospectra are dominated by arboreal (*Pinus, Betula*) and herbaceous–fruticose (*Artemisia*, Graminaceae, Cyperaceae) pollen. The herbaceous group is characterized, similarly to the previous section, by the composition corresponding to plant communities of the meadow–forest (Ericales, Onograceae, Plantaginaceae, *Ranunculus, Cacalia hastata, Heracleum sibiricum*) and steppe–meadow (*Lilium tenuifolium, Centiana barbata, Pulsatilla patens, P. bungeana, Bunias, Thymus, Ephedra*) types. Pollen of broad-leaved forms (*Ulmus, Tilia, Acer, Yuglans*) is also present.

The RTL date for the Khidusa section, the last one of a series of the middle Pleistocene sections, is 156 ± 39 Ka. The share of the arboreal–fruticose pollen in the palynospectra from this section exceeds 90% and mainly consists of abundant specimens of *Pinus sylvestris* and *Betula*. Broad-leaved trees are represented by *Ulmus* and *Tilia*.

Thus, deposits that were formed, according the RTL dating, in the time interval from 390 ± 80 to 156 ± 39 Ka occur within the Baikal-oriented drainage at a significant height above the recent deposits of the Anga, Bugul'deika, and Goloustnaya rivers. Data on sediment accumulation in valleys are yet insufficient to consider the sedimentation history in detail and we regard this interval as the undivided middle Pleistocene sedimentation stage.

The upper Pleistocene horizon of the Manzurka alluvium is studied in valleys of the Bugul'deika, Ungura, Manzurka, and Lena rivers, where deposits

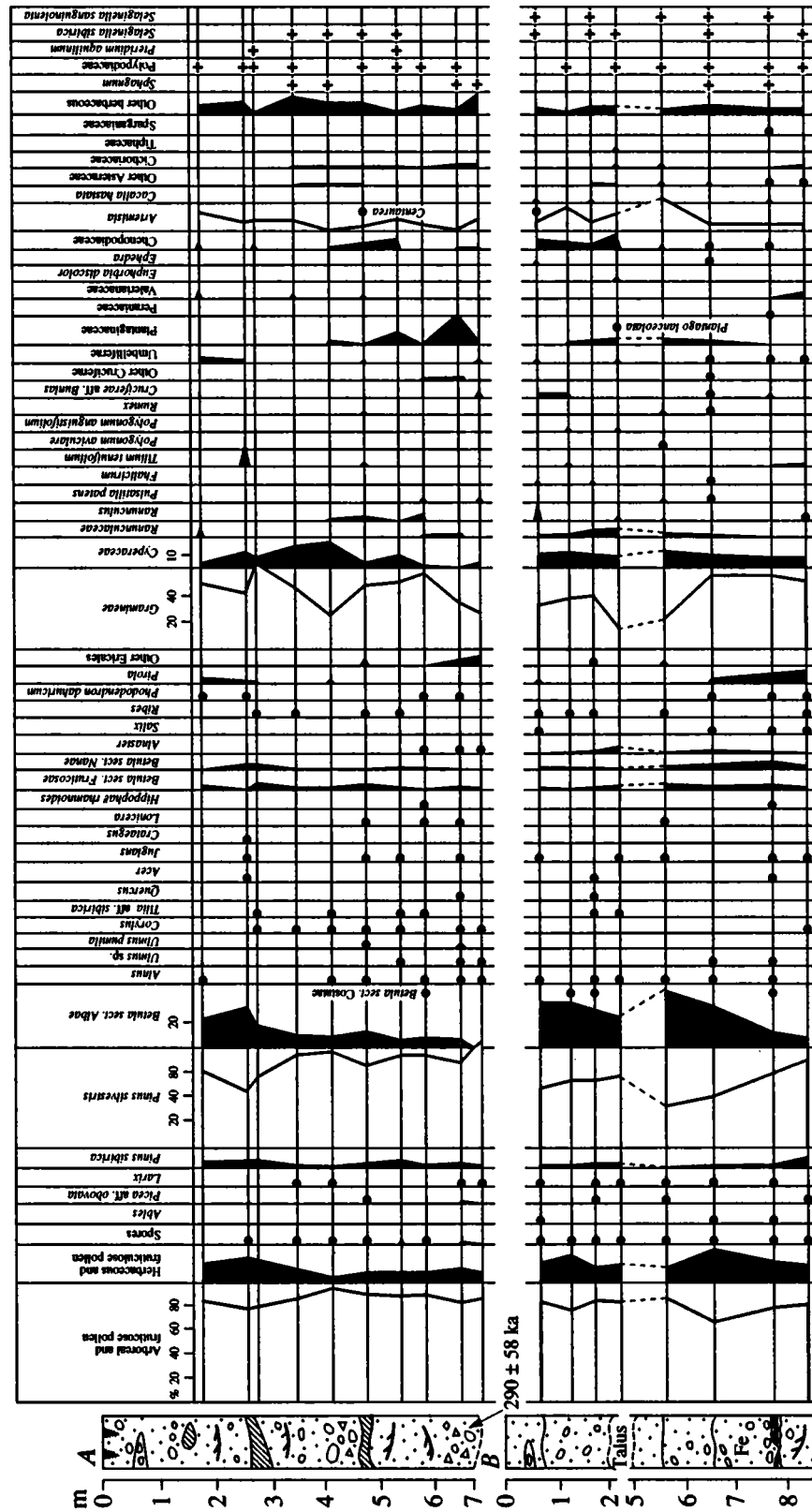

Fig. 3. Spore–pollen diagram for middle Pleistocene deposits of the Elantsy (A) and Bugul'deika (B) sections (symbols as in Fig. 2).

show close RTL dates. All sections are located near the valley sides and often compose terrace-like elevations in areas of the valley widening. In the Bugul'deika River valley, these deposits are exposed north of the highway between Kosaya Step' and Elantsy settlements and stretch to the Bilet Creek inflow. There are several sections located along the distance over 12 km along the left river side; the Southern section is studied most thoroughly.

Palynospectra from the Southern section yield 19 genera of arboreal vegetation representing up to 90% of the total pollen assemblage. The pollen of *Pinus sylvestris* is particularly abundant. The content of herbaceous pollen is 10% and less; spores are encountered as single grains. Elements of meadow–forest (*Rannunculus*, Ericales, Onograceae) and meadow–steppe (*Lilium tenuifolium, Rheum, Tanacetum, Allium, Ephedra*) vegetation were determined in addition to representatives of widespread families (Gramineae, Cyperaceae) and genera (*Artemisia, Polygonum*). According to the RTL dating, the age of deposits is 133 ± 30 Ka.

In the Ungura River valley, the upper Pleistocene deposits (RTL dates are 140 ± 35 and 108 ± 27 Ka) are recovered in the quarry located in southern outskirts of the Kaptsygai Settlement. Palynospectra are dominated by the arboreal pollen (more than 90%) represented by 15 genera (90% of *Pinus sylvestris* and not more than 10% of *Betula*). Single pollen grains of the broad-leaved flora are recorded in some samples.

The more complete palynological characteristics were obtained for the 20-m-thick sequence north of the Nikilei Settlement, where these deposits cover a surface of the terrace-like elevation about 35 m high. There, the section is composed of alternating thin ferruginous pebbly and sandy–gravelly beds displaying cross bedding. Palynospectra from the section contain 19 genera of the arboreal vegetation amounting to 74–94% of the total pollen abundance (Fig. 4). In the upper part of the sequence, all floral groups reveal changeable relative proportions. The share of herbaceous–fruticose polen decreases. Simultaneously, gramineous forms, *Artemisia*, and representatives of Compositae become less abundant in general. In the arboreal pollen assemblage, the share of *Pinus sibirica* increases whereas that of *Betula* sec. Albae decreases. It can be suggested that these changes are not incidental but reflect a real growth of the distribution area of dark coniferous formations with *Pinus sibirica* in response to the general elevated humidity in the region. It is probable that these changes correspond to the initiation of the period of moistening and cooling; this assumption is also supported by changes in the herbaceous community. Nonetheless, the permanent presence of pollen of broad-leaved forms in palynospectra does not suggest a significant cooling. Based on the available data, only a weak cooling tendency can be noted. The RTL date is 115 ± 30 Ka.

Two sections were studied in the Lena River valley at the left side. In the quarry located here in the southern outskirts of the Kachug Settlement, the coarse clastic sequence overlying the Cambrian Verkhnyaya Lena Formation is exposed at the height of about 45–50 m above the water line. Palynospectra are dominated by pollen of arboreal–fruticose vegetation, including abundant pollen of *Pinus sylvestris*. Dark coniferous forms make up 4–7%; a significant share of the spectrum belongs to arboreal and fruticose forms of the *Betula* genus. Pollen of *Ulmus, Corylus, Tilia, Quercus, Tilia*, and *Caprinus* is also recorded. The RTL date is 126 ± 25.

Another section was studied in the quarry located at the right bank of the Lena River, east of the Kistenevo Village. The thickness of the sequence is 40–42 m. Its upper part is mainly composed of sandy–pebbly–gravelly deposits with small loam lenses. Rocks of this part show the lenticular and horizontal bedding, whereas the lower part is represented by pebbly layers with sandy lenses and interbeds.

The pollen assemblage is diverse. The arboreal pollen represents 23 genera, most abundant of which are *Pinus sylvestris* and *Betula* (Fig. 5). Palynospectra of most samples contain pollen of several broad-leaved taxa, particularly that of *Tilia* and *Ulmus*. The herbaceous group includes more than 60 species including many representatives characteristic of steppe formations and pioneer groupings populating trains, river bed shingle, and fans (*Allium, Convulvulus, Taraxacum, Filifolium, Thaesium, Tanacetum*, Malvaceae, Centaurea). Spectra correspond to dark coniferous, mixed, and steppe forests as well as to meadow–steppe and petrophitic steppe groupings. Mixed forests and valley groves included elements of broad-leaved flora. The RTL date is 78 ± 20 Ka.

Peculiarities of Pleistocene palynospectra

Palynospectra for the middle and first half of the upper Pleistocene are close in composition. They are dominated by arboreal pollen with an insignificant admixture of spores. The prevailing element is represented by *Pinus sylvestris*, and pollen of *Larix*, one of the main forest-forming forms, occurs as single grains. The deposits also contain pollen of broad-leaved flora. The herbaceous–fruticose group is characteristic of mixed forest, meadow, and steppe formations.

The flora and plant communities of the western Baikal region were characterized by similar features during all studied periods. The inferred uniformity of distribution patterns of vegetable formations in low mountains and elevated plains seems unusual. The region under consideration appears to be covered by the southern coniferous taiga and by forest–steppe pine–birch and birch forests, as well as by herbaceous meadow steppes. In addition, several species of the broad-leaved flora were constituents of zonal forests

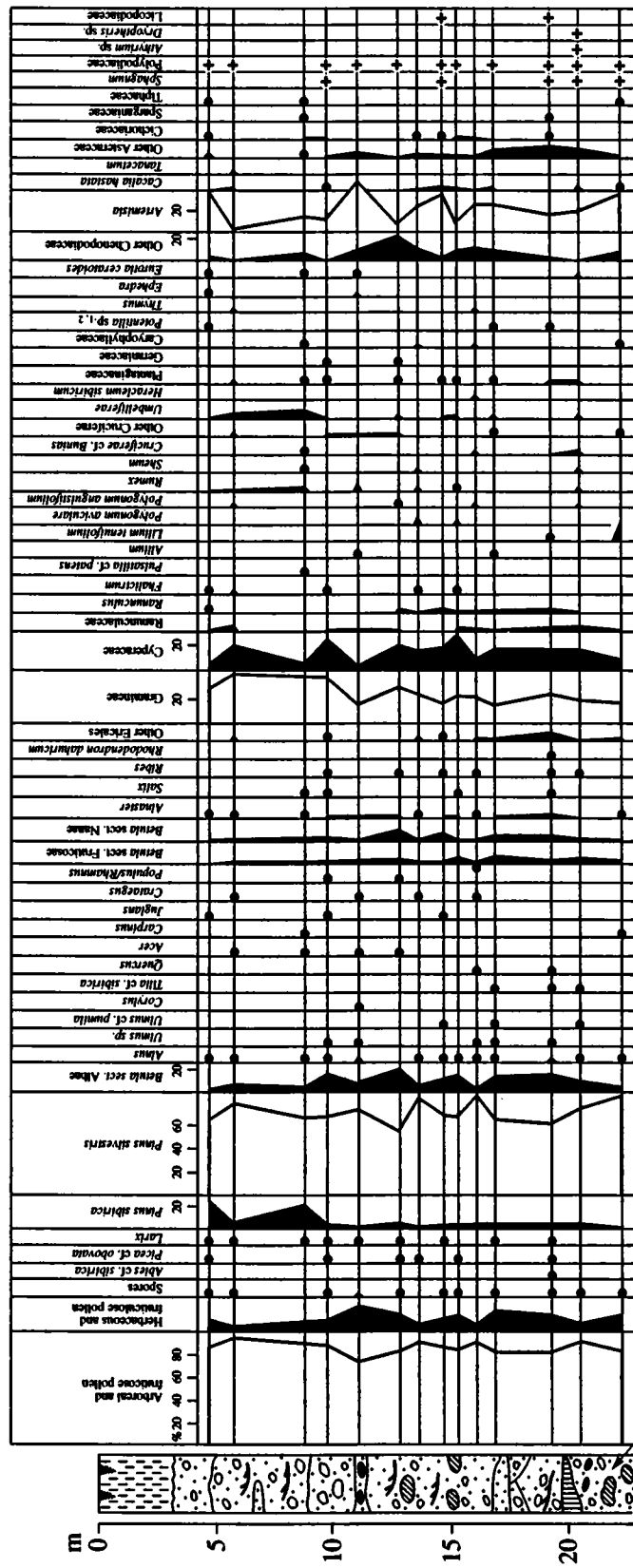

Fig. 4. Spore–pollen diagram for upper Pleistocene deposits of the Nikolei section (symbols as in Fig. 2).

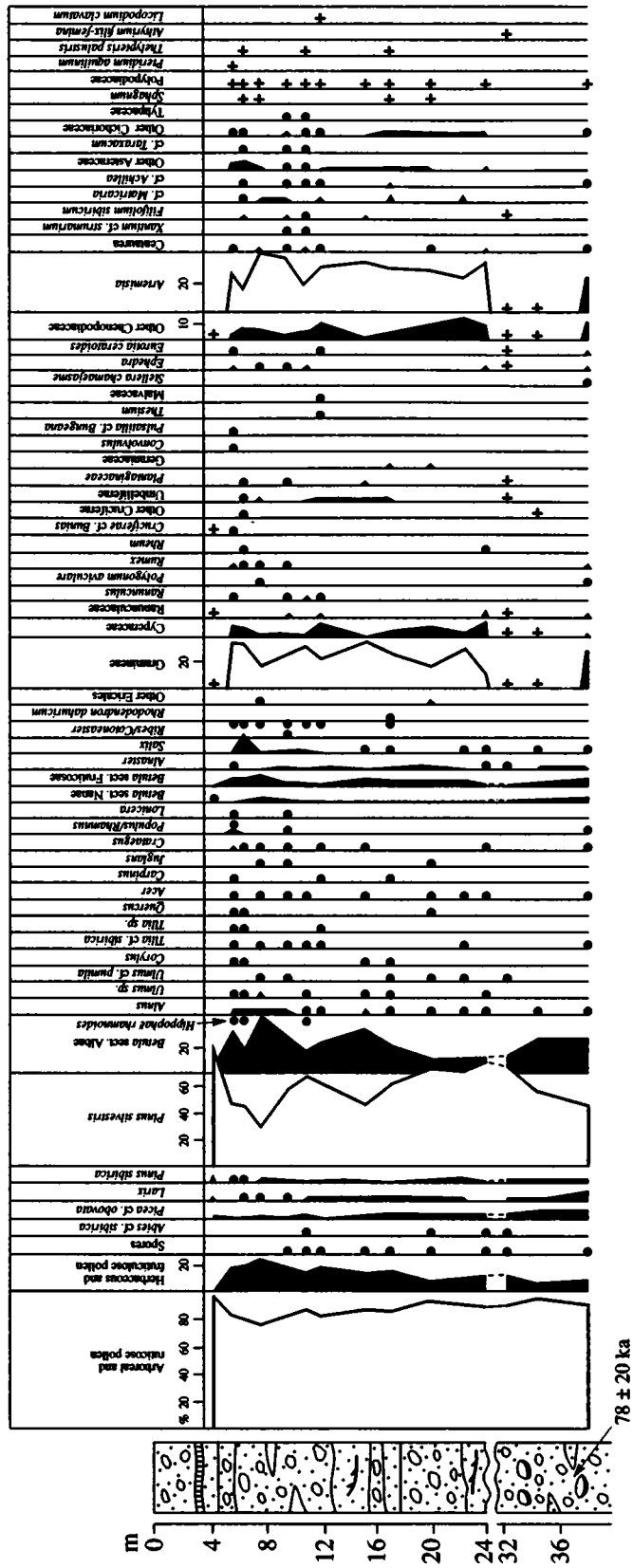

Fig. 5. Spore–pollen diagram for Upper Pleistocene deposits of the Kistenevo section (symbols as in Fig. 2).

and valley vegetation thus suggesting the development of mixed broad-leaved formations representing the zonal forests. The role of *Larix* in the vegetation cover was significantly lower than nowadays. This genus became a dominant element of mountainous forests in the Baikal region in the late Pleistocene.

Palynological data do not indicate significant changes in the vegetable cover, although they are incomplete, and the problems of detailed climatostratigraphy are not discussed in the present work. Nonetheless, the pollen of broad-leaved flora established in deposits of all studied stratigraphic intervals including sediments, which are referred on the basis of the RTL dating to the Samarovo and Taz stages, is of great interest.

Evolutionary Stages of Vegetation and Sedimentation in Valleys of the Western Baikal Region

The formation of constrative sequences of the Manzurka alluvium in valleys of the western Baikal region continued (with breaks) for a long period. This period includes three major stages of flora and vegetation evolution. The first stage corresponds to the development of polydominant coniferous–broad-leaved forests with an admixture of the diverse group of exotics (N_1^2).

The second stage was characterized by the impoverished Pliocene flora, which experienced an episode (episodes?) of aridization and was transformed into a peculiar floral assemblage that included elements of South European–Asian nemoral forests (*Fagus, Polystichum lobatum*) and Mediterranean–Asian steppe vegetation (*Elaeaghus, Hippopha rhamoides, Eremurus*). Forest formations were dominated by modern forest-forming forms (*Pinus sylvestris, P. sibirica, Picea, Abies, Betula*). The group of broad-leaved, coniferous, and other exotics was barren of elements from the American–East Asian (*Tsuga, Diervilla*) and American–Eurasian (*Fraxinus, Ilex*) distribution areas. During this late Pliocene stage, the mixed broad-leaved forests represented one of the zoned types of vegetable formations.

The third stage corresponds to the late Pliocene and Pleistocene and is characterized by the prevalence of southern taiga and steppe forests with an admixture of the broad-leaved flora (mainly in valley forests and in those growing near the valleys).

Transformation of the flora and vegetation indicates a certain tendency of climatic changes in the region. The formation of constrative alluvium continued in the Late Cenozoic during a period of about 4 m.y. regardless of the climatic situation. The role of climate in the process of constrative alluvium sedimentation in valleys was apparently insignificant.

The available data allow six pulses of sedimentation to be defined (table), though their number was probably greater.

DISCUSSION

We should answer several questions, with the first of them being the following: what is the Manzurka alluvium? In typical sections, for instance, in the quarry near the Kosaya Step' Settlement on the right side of the Bugul'deika River, it is a constrative alluvium composed of alternating layers and lenses of gravelly to pebbly sands displaying cross bedding. The clastic material is virtually unsorted throughout the section, though some sorting is notable within cross-bedded layers. The alluvium beds proper are almost identical in composition, and the bedding proper is often obscure. Another lithological feature of the Manzurka alluvium is a distinct preponderance of the river bed facies in its sections; oxbow faces are rare.

In most sections, the Manzurka alluvium is light-gray (whitish), and its clastic material is slightly weathered. The light-gray coloration is usually combined with the films, spots, or bed-confined brown to red–brown aggregates of Fe hydroxides responsible for variegated outlook of the rocks.

The inner structure of the Manzurka alluvium is determined by the occurrence of members with oblique bedding against the background of uniform lithology that is characteristic of constrative alluvium. Another peculiar feature of the alluvium is its localization in ancient valleys, which are only partly used by current rivers. This peculiarity of the Manzurka alluvium is tipical of all its stratigraphic units ranging from Neogene to Pleistocene in age. The ancient valley in the interfluve area between the Kuyada and Bugul'deika rivers, whose bottom is located at the altitude exceeding 100 m above the Lake Baikal, and another valley in the direct proximity to the lake (they are separated only by narrow coastal ridge) are filled with the Pleistocene alluvium representing a good example of deposits under discussion.

Despite the visual lithological uniformity of the Manzurka alluvium, it includes units different in age. This feature of the Manzurka alluvium is established first after our study and simultaneously raises several question concerning the structural and paleogeographic aspects of its formation.

The next question is to what extent the Manzurka alluvium is an incidental type of deposit? The answer is that it is not unique. The so-called whitish alluvium of the eastern Baikal region, whose age was once debated widely (Kornutova and Khotina, 1967; Simonov, 1972; Malaeva and Ufimtsev, 1976; and others), provides another example of similar sediments. This is a thick (100 m and more) sequence of cross–bedded light-gray gravelly sands occurring either on sides of modern and in ancient valleys, or filling shallow Cenozoic depression. In lithology and genesis, the Manzurka alluvium of the Baikal region and whitish alluvium of the Transbaikal region are completely similar: both of them represent sediments of large rivers, which longly transported the material. The whitish alluvium of the eastern

Main formation stages of the Upper Cenozoic Manzurka deposits in the Baikal region

Geological age	Radiological dates, Ka (RTL)	Regional horizon	Sedimentation basins, genesis of deposits	Neotectonic events	Sedimentation stages
Q_3^{2-3}			Mainly river incisions	Tectonic scarp formation in the Primorskii Ridge	
Q_3^{1-2}	78 ± 20 108 ± 27 140 ± 35 115 ± 30 126 ± 25 133 ± 30	Nikelei	Alluvium of medium-high terraces-like spurs in valleys of the Lena, Bugul'deika, and Manzurka rivers		6
			Local reorganizations in the drainage system	Differentiated movements of tectonic blocks in the Baikal region	
Q_2	156 ± 39 230 ± 60 290 ± 58 390 ± 80	Khogot	Alluvium of terraces (Khodontsa R.); alluvium of the Baikal region buried in modern valleys of the Goloustnaya, Bugul'deika, and Kuyada rivers		5
			Deposits are not recorded	?	
		Podtok	Alluvium of the high level (80–90 m) in the Manzurka River valley (Kharbatovo sections)		4
			Mainly river incisions	?	
N_2^2		Kharbatovo	Alluvium of the high level (80–90 m) in the Manzurka River valley (Kharbatovo sections)		3
			?		
		Anga	Alluvium in the Anga River valley (tributary of the Lena River)	Relative subsidence (?)	2
			Deposits are not recorded	?	
N_2^1		Bugul'deika	Alluvium in local depressions		1

Baikal region represents polychronous deposits that consist of several sequences different in age (Simonov *et al.*, 1988). When all the peculiarities of loose deposits from adjacent regions are taken into consideration, it is clear that polychronous (age-variable) lithological complexes represent a characteristic phenomenon of the geological structure in the southern part of East Siberia. The Krivoi Yar sands of the western Transbaikal region (Ravskii *et al.*, 1964) that include sediments of several age generations (Bazarov, 1968) are of a similar origin. In the upper part of the Cenozoic section in the Tunka rift valley, three polychronous continuous sedimentary complexes can be defined: sandy, blanket-forming, and boulder–pebbly. The latter represents block–boulder–pebbly deposits of river fans, whose total thickness is up to hundreds of meters. The lower part of the complex is composed of boulders, shingle, and conglomerates of the Pliocene Anosovo Formation (Logachev, 1958, 1968).

New data on the structure and age of the Manzurka alluvium raise several problems of the Cenozoic geology in southern East Siberia. Now, we should speak not about this stage but about Manzurka stages, because it is clear that this alluvium includes several units different in age. This means that either a single system of drainage from Baikal toward the Lena River basin existed in the western Baikal region during the long time period or this situation repeated intermittently. Of course, the problem of relationship of rivers and ancient valleys that existed at the "Manzurka time" around Lake Baikal is yet to be solved.

The lithological uniformity of the Manzurka alluvium sequences suggests some stability of physical–geographical environments during their formation. A strong preponderance of a sandy-gravelly material indi-

cates a common type of bed rock weathering throughout the Pliocene and Pleistocene—mainly saprolitization and the formation of large amounts of grus (Simonov, 1972). This indicates the progressive and relatively stable climatic conditions at that time. The progressing successive transformations of vegetation communities are inferred from palynological characteristics of the Manzurka alluvium.

Structural problems concerning the Manzurka alluvium and similar polychronous continuous sequences of loose deposits appear to be complex. First of all, there is a problem of stratigraphy and correlation of lithologically uniform sequences with a complex inner structure. Reliable lithological and biostratigraphic chronometers should be elucidated and methodical approaches should be elaborated for determining and describing spatial relationships of sequences comprising units of a different age.

Finally, there are also theoretical problems. The Manzurka alluvium is, in fact, an analogue of similar pre-Cenozoic deposits. From the viewpoint of study and mapping of the latter, the Manzurka alluvium represents a perfect model object of polychronous continuous complexes of loose deposits. They are widespread in the inner areas of Eurasia being formed under stable or gradually changing physical–geographical conditions in the intracontinental regions. In this situation, the concept of glacial and interglacial epochs becomes inadequate, and lithological or biostratigraphic chronometers for loose deposits should be specific. The history of formation of continuous complexes of loose deposits is an extremely interesting geological problem, and the Manzurka alluvium represents an ideal object for such investigations.

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A Discussion of Problems of Precambrian Geology and Stratigraphy at the Third All-Ukraine Interdepartmental Meeting, April 1998

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The Third All-Ukraine Interdepartmental Meeting on *Precambrian Geology and Stratigraphy of the Ukrainian Shield* was held in Kiev, April 21–23, 1998. The purpose of the meeting was to analyze new data on the Precambrian stratigraphy and geochronology of the shield, obtained since the last meeting in 1991. Representatives from different scientific institutions, industrial enterprises and universities of the Ukrainian and Russian Academies of Sciences presented and discussed 28 oral communications and 25 posters. The theses of all papers were published in Russian and Ukrainian in *Geologiya i stratigraphiya Ukrainskogo schita. Tezisy Vseukrainskogo mezhvedomstvennogo soveshchaniya. Kiev, aprel' 1998* (Geology and Stratigraphy of the Ukrainian Shield. Theses of Papers of the All-Ukraine Interdepartmental Meeting. Kiev, April 1998). The authors of the papers treated practically all aspects of geology of the Ukrainian Shield (USH) and analyzed the results of geological mapping, prospecting data on useful minerals, and isotope dating of rocks and minerals.

Three papers were received with particular interest. N.P. Shcherbak presented a review of the most important problems of Precambrian geology and stratigraphy and gave an analysis of the state of isotope–geochronological studies of the main subdivisions of the Precambrian formations in the shield. He also outlined their correlation with the units of the International Precambrian Scale. D.S. Gurskii emphasized the priority of Precambrian geology in the further development and improvement of the mineral and raw materials base of Ukraine. K.E. Esipchuk characterized the stratigraphic problems specific for the Precambrian, and treated in detail the principles of compilation of the regional stratigraphic scale from the standpoint of peculiarities of the geological structure of the shield.

The communication by E.B. Glevasskii, Esipchuk and Shcherbak analyzed the important geological–geophysical and petrological–geochemical aspects of structure and evolution of the USH lithosphere. The concrete problems of correlation of the USH Precambrian were discussed by G.V. Artemenko, Shcherbak and L.M. Stepanyuk with regard to the Archean and by Glevasskii, V.M. Skobeleva and Stepanyuk with regard to the Proterozoic. E.M. Laz'ko and co-authors demonstrated possibilities of the geological–formational approach to the cartographic systematization of geological knowledge; they also analyzed the ensuing results of such an approach and original concept of principal importance for the understanding of the peculiarities of successive events in the Early Precambrian history of the Earth. V.V. Zyul'tse dedicated his report to the western part of the USH and pointed out the problem principally important for geological practice and adequate understanding of geological history from the information presented in geological maps. Zyul'tse also emphasized the primary importance of geological mapping and elaboration of its stratigraphic basis. Proceeding from the results of the geological–reconnaissance studies of 1991–1997 in the Sea of Azov region, V.F. Razdorozhnyi suggested a stratigraphic scheme, which includes the Kayinkul', Upper Tokmak and Central Priazovskaya groups of the Lower, Middle and Upper Archean, respectively, the Lower Proterozoic Orekhov Group, the Middle Proterozoic Osipenko and Gulyai Pole groups, and the Kolarov, Kil'tich'ev and Zelenaya magmatic complexes of the Upper Proterozoic. V.P. Kirilyuk demonstrated, using the example of the western part of the USH, the basic importance of the geological–petrological data for understanding the stratigraphy of deeply metamorphosed complexes.

V.Z. Negrutsa from the Geological Institute of the Kola Scientific Center, Russian Academy of Sciences, informed the participants that the Permanent Commission for the Lower Precambrian of the Interdepartmental Commission on Stratigraphy (ICS) of the Russian Federation is now elaborating the network concept of reference sections and boundary standards for the Lower Precambrian stratigraphic units; he attributed particular importance to the general significance of work coordination in this field within the Baltic and Ukrainian shields and in the Voronezh massif. V.M. Klochkov and his colleagues Ya.P. Bilinskaya, G.P. Khvorova and Yu.K. Piyar described the methods and main results of organization of database for the Precambrian reference sections of the USH.

Several presentations were dedicated to new data obtained in the course of geological surveys carried out during the last few years. V.V. Sukach and co-authors have established that an independent Serga complex of subvolcanic formations comagmatic with the Apollonia basaltoids is distinctly isolated from the Sura complex of the Sura greenstone trough. V.V. Zakharov, V.K. Butyrin and N.S. Kurlov presented the geological survey results on the area of M-36-XXXIV and L-36-IV map sheets and summed up the information obtained from the Krivoi Rog superdeep borehole and supplementary wells. The authors demonstrated the reference significance and synchronism of the black carbonate schist levels in the Rodionovka and Gdantsevo formations and their importance for correlation of the Krivoi Rog Group with the lower part of the Ingul–Ingulets Group. According to S.P. Lashko, the results of geological–prospecting studies of the last decade imply the possibility of obtaining a unified scheme of layer-by-layer subdivision of the iron ore deposits of the Krivoi Rog and Ingulets groups. These results indicate the identical age of rythms of the low and highly ferriferous units and synchronism of the processes of iron accumulation over the entire area of the Krivoi Rog–Kremenchug paleobasin. According to the data presented by N.Ya. Yashchenko and V.V. Shekhotkin for the region of the Klintsov and Yur'ev gold-bearing deposits, the rocks of the Checheleevka Formation are pierced by granites and gabbro–diorites of the Late Archean (3100–2750 Ma), whereas in the scheme by Zakharov and co-authors, the Checheleevka Formation is attributed to the upper part of the Ingul–Ingulets Formation. According to the report presented by R.Ya. Belevtsev, "the ferruginous quartzites of the USH are not sedimentary, but of intrusive–magmatic liquation genesis, and represent the Archean rocks formed as a result of the intrusion of ferruginous–siliceous magmas into the ancient tonalite–trondhjemite complex of gray gneisses."

Data important for better determination of the historical–geological basis of the Precambrian periods were presented in the generalizing reports by G.I. Kalyaev *et al.*, L.S. Galetskii *et al.*, Glevasskii and Kalyaev, Yu.P. Orovetskii *et al.*, and E.A. Kulish. The authors considered problems of tectonic structure and evolution of the Early Precambrian crust on the USH and made comprehensive analyses of the geological, geophysical and isotope-geochemical data with regard to the metallogeny of the area. V.A. Ryabenko and T.P. Mikhnitskaya drew attention to the fact that, after the formation of the Korosten' rapakivi granite complex (1750 Ma ago), the USH experienced several transformation stages correlative with the Vyborgian (1600–1500 Ma ago), Kibarian (1400–1300 Ma ago) and Grenvillian (1100–950 Ma ago) orogeneses. Mikhnitskaya dedicated her paper to the general regularities in the riftogenesis and sedimentation of that period in the territory of the USH.

Considerable attention at the meeting was drawn to the endogenic and, primarily, magmatogenic aspects of the USH geology. The report by I.B. Shcherbakov *et al.* outlined the structure of the Vorovskogo massif of the ultramafic–mafic–monzonite rock series (1970–2010 Ma), which they compared with the Elan'–Vyazovka pluton of the Voronezh massif. These authors also dwelled on the structure of the Bushveld, Sudbury and Stillwater plutons. V.P. Bezvinnyi and V.I. Orsa, who used the results of geological survey at the scale 1 : 200000 in the territory of the Uman' and Belaya Tserkov' areas, established three periods of granite formation and showed their age to be 2061 ± 15, 2030 ± 5, and 1983±12 Ma. Razdorozhnyi *et al.* presented the results of study of nonmetamorphosed micaceous ultramafites, one body of which (Mariya Pipe) intrudes diorites of the Kolarov massif about 2000 Ma old and is pierced by the veins of plagiogranites 1760 Ma old. S.G. Krivdik showed that alkaline rocks of the USH, unlike similar rocks in other shields, are Proterozoic in age and have been formed in two stages: 2100–1900 and 1800–1700 Ma ago. N.N. Kir'yanov, V.Yu. Chernov *et al.* characterized the kimberlites of the Kirovograd geoblock, where they are of three age groups: (a) 1700 Ma old; (b) 1500–1200 Ma old, and (c) 1200 Ma old. L.V. Shumlyanskii, O.V. Zinchenko and E.G. Molyavko dedicated their paper to the trap formation in the south of the East European platform, which originated 1900–2000 Ma ago. Krivdik *et al.* showed that the eastern part of the Sea of Azov massif experienced 500–250 Ma ago the tectono–magmatic activation. Attention was also drawn to the well-known spores and vegetation remains in the Precambrian metasedimentary rocks of the USH. A.M. Snezhko repeated her earlier inference that these remains are allochthonous, redeposited in the course of formation of the Phanerozoic weathering crusts.

The new correlation scheme of the Precambrian formations in the USH was the general outcome of the meeting. This scheme was approved at a joint session with the Archean and Paleo-Mesoproterozoic Commissions of the National Committee on Stratigraphy of Ukraine (NCSU). The scheme was also recommended for authorization by the Third All-Ukraine NCSU Meeting that was done on May 24, 1998.

The main part of the scheme includes the regional stratigraphic units and complexes of endogenic formations (ultrametamorphogenic, magmatic). Six regions, different in structure and composition of the Precambrian stratisphere, are deliniated, and each of them is characterized by two individual columns: the stratigraphic and the metamorphic–magmatogenic, placed respectively in the left and right-hand parts of the scheme. Thus, the left column shows correlation of regional lithostratigraphic subdivisions (groups, formations, and units) in their actually observed succession, whereas the right one represents the correlation of chronological successions of magmatic and metamophic complexes, among which the autochthonous,

para-autochthonous and intrusive subdivisions are shown separately. The natural boundaries, which divide the sequences into subordinate complexes, are represented by structural unconformities and other boundaries between events detectable on the basis of combination of geological and isotope–geochronological data. Five boundaries are distinguished on the USH territory, where the Precambrian is divided into six stratigraphic high-rank subdivisions: the *Dniester Supergroup* (older than 3400 Ma) composed of the Dnieper–Bug, Aul' and Western Azov groups; the *Azov Supergroup* (3400–3200 Ma) of the Bazavluk and Kosovitsa units; the *Dnieper Supergroup* (3200–2600 Ma) of the Bug, Ros'–Tikichi, Konka, Central Azov groups and their analogs; the *Krivoi Rog Supergroup* (2600–2000 Ma) composed of the Teterev, Ingul–Ingulets, Gleevataya and former Krivoi Rog groups of the Dniester region, and also the Gulyai Pole, Dibrovka formations and Sadovoe Group of the Sea of Azov area; the *Klesovo Supergroup* (2000 –1700 Ma) represented by the Klesovo Group, Pugachev Beds and the Topilyansk Group of the northwestern part of the USH; and the *Terminal Supergroup* (1700–1500 Ma) actually corresponding to the Ovruch Group. The interregional stratigraphic complexes universal for the entire USH are regarded as regional representatives of the general subdivisions of geological time: the eo-, paleo-, meso- and neo-Archean erathems of the Archean acrothem, and the paleo-, meso-, and neo-Proterozoic eonothems of the Proterozoic acrothem. The boundary ages of the general subdivisions of geologic time, which are approved by the International Subcommission on Precambrian Stratigraphy of the International Union of Geological Sciences (IUGS), are arbitrarily changed in accordance with the isotopic dating results for the USH complexes. Concurrently, these dates are used to calibrate the regional complexes in accordance with the linear time scale at the resolution level of 100 m.y. In the course of discussion of all the problems mentioned above, the participants repeatedly emphasized the insufficient accuracy of geohistorical determinations of the boundaries between Precambrian regional stratigraphic subdivisions of the USH and the necessity for more exact estimation of both the subdivision ranges and isotopic ages of their boundaries.

The meeting results and analysis of the available materials used as a basis for discussion reveal five points which, as we believe, are of general importance for the future improvement of methodology and studies of the Precambrian stratigraphy of metamorphosed formations.

(1) The detailed and reliable stratigraphic subdivision of Precambrian deposits is a direct function of the preservation degree of original lithologic, petrographic and facies–paleogeographical properties of stratified rocks.

(2) The magmatic–metamorphic changes in the Precambrian deposits disguise the chronological indicators in succession of accumulation stages of stratified rocks, which are basic in the stratigraphic chronology, into thermobarometric characteristics of endogenesis changeable with time and over an area, and also depending on the geodynamic settings and tectonic history of a concrete crustal element. Thus, the isotopic ages of metamorphic complexes reflects the thermobarometric history of stratigraphic units but not the time of their formation, and, therefore, they are of indirect significance for subdivision and correlation of the Precambrian sections.

(3) The actually observed (mapped) local stratigraphic (structural–lithologic or formational) subdivisions and their boundaries reflect the irreversible course and natural boundaries of events in the geological history. Moreover, the geological time is recorded in these subdivisions depending on the completeness of a given section, and interpretation of recorded events is inevitably subjective.

(4) Correct stratigraphic inferences should be based on results of geological mapping. Because of the current reduction of geological survey, the effectivity of stratigraphic studies depends on the choice of investigation methods and on the quality and completeness of geohistorical information for an object chosen as a standard.

(5) Most important for improvement of the Precambrian regional and general stratigraphic scales are the choice of stratotypes (the type sections and boundary points of stratigraphic subdivisions), investigation of their genetic history in order to broad the available set of iterrelated correlation criteria, and dating of events recorded in the sections.

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